



UNIT 27

EARTH MATERIALS AND PROCESSES: AN INTRODUCTION TO THE ROCK CYCLE

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STUDY GUIDE

In Units 27 to 29 we return to the Earth sciences, first introduced in Units 5 to 8, and focus in a little more detail on the materials of which the Earth is made and their history. Unit 27 consists of four components: the text, an experiment, two AV sequences and a TV programme. While studying it you will need the rock and mineral specimens and the hand lens from the Kit. For the first AV sequence you will also need dilute hydrochloric acid, the watch glass and a pipette, and for the experiment you will also need the balance and a 100 cm³ beaker. Early in Section 2, the first AV sequence 'Minerals' (Tape 4, Side 1, Band 2) will help you examine the mineral specimens in some detail, and later in the Section there is an experiment to measure the density of mineral and rock samples. For the experiment you will need to provide something with which to suspend the rock samples: fuse wire is best, and strong cotton will do, but you may find a piece of the netting container from, for example, oranges, onions or garlic, or an old hairnet less fiddly! In Section 5 the third major group of rocks, the metamorphic rocks, is introduced and the second AV sequence 'Metamorphic rocks' (Tape 4, Side 2, Band 1) explains their identification and characteristics. The TV programme 'From Snowdon to the sea' may be watched with equal benefit at any time in your study of the Unit, but is particularly relevant to Sections 4 and 5.5. The notes for the programme are in Section 4.8.

I INTRODUCTION

In the earlier Earth sciences Units you learned about the gross physical and chemical structure of the Earth, how this is modelled and the advances in our understanding of how the world works that were brought about by plate tectonic theory. Now, having learned a little about the biology of the natural world, we return to the physical natural world, to the processes that make and shape the continents on which we live. Some of the questions we answer in this Unit are:

Why are the volcanoes of the Andes more dangerous than those of Hawaii?

How are sediments formed?

What are the processes involved in turning mud into slate?

Are the processes involved in the questions above linked?

At first you may find it difficult to appreciate the timescales involved in geological processes. Some famous rock outcrops from around the British Isles—the Pembrokeshire coast, Arthur's Seat, Giant's Causeway and Land's End—are illustrated in Plates 1 to 4 (at the back of this binding). In these Plates, rocks appear static and unchanging; this is because many geological processes take place on timescales of thousands to millions of years, which are quite different from our own. Units 28–29 cover some of the aspects of geological time and its measurement, and some of the milestones in the evolution of our planet and its lifeforms during Earth history.

In this Unit, we take up the theme introduced in Units 5–6, that rocks are being formed all around us, all the time. But geological timescales are very long compared with our lifetimes; if we lived to an age of 70 000 years, rather than 70, it would only be a little easier to appreciate the ways in which different rocks can be made from a common starting point!

The starting point for this Unit is peridotite. As you will recall from Units 5–6, different types of peridotite form the Earth's mantle, i.e. 80% by volume of the Earth. As you will see later in this Unit, it is thought that almost all crustal rocks are derived either directly or via a series of complex indirect processes from mantle peridotite. Your sandstone specimen S6 doesn't look anything like the peridotite S4 and, chemically, it is totally dissimilar, but it was derived from peridotite. The derivation of the one

HYDROLOGICAL CYCLE

ROCK CYCLE

ROCK

MINERAL

rock from the other is usually not a single-stage process: in most cases it requires complex repetitions of similar processes that form part of the Earth's *geological cycles*.

1.1 GEOLOGICAL CYCLES

Geological cycles act within the Earth on large and small scales, and over long and short periods. They may affect large masses of rock, or individual elements; they may be simple or complex, but they all involve the manufacture and re-processing of the Earth's materials.

One important geological cycle is the water cycle, or **hydrological cycle**. In this cycle water, in the form of rain, hail or snow, that falls on land is carried to the sea in rivers, and then evaporates, condenses, and falls again. This is illustrated in Figure 1.

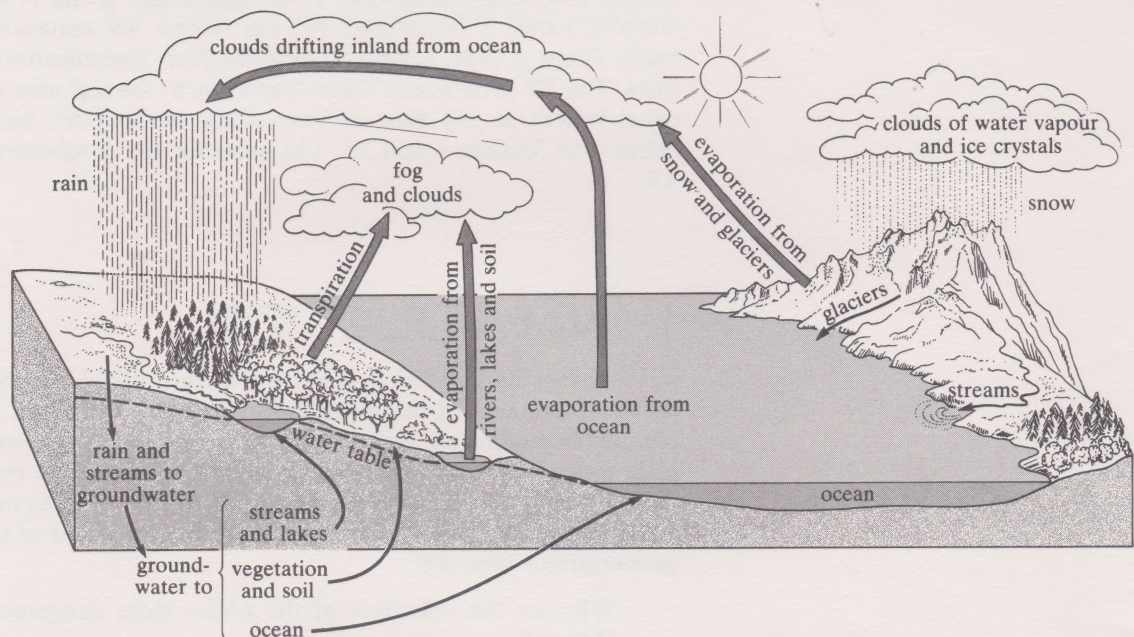


FIGURE 1 The hydrological cycle. Movement into the atmosphere by evaporation is matched by precipitation as rain and snow.

Intimately linked to the hydrological cycle is the best-known cycle in geology, the **rock cycle**. This was first recognized by the pioneering Scottish geologist, James Hutton. In a book published in 1785 called, rather immodestly, *The Theory of the Earth with Proof and Illustrations*, he showed how:

- 1 igneous rocks may be eroded to form sediments by weathering and erosion;
- 2 the sediments may become compacted into rocks; and
- 3 a later mountain-building event may then expose these sedimentary rocks at the Earth's surface, where they may be eroded away, thus forming a fresh generation of sediments.

The parts of the cycle are illustrated in Figure 2.

A cycle of any kind requires an energy source to drive it. What are the energy sources driving the hydrological and rock cycles?

The energy source driving the hydrological cycle is relatively easily identified as the heat of the Sun. The rock cycle is more complex; the hydrological cycle is involved because moving water plays a major part in the formation of sedimentary rocks. Equally important, however, is the Earth's interior heat, which is responsible for driving plate tectonic processes and for building up the mountain belts, which are subsequently eroded. The main theme of this Unit is to consider the components of the rock cycle, section by section. First, though, we must look more closely at the materials of which rocks are made.

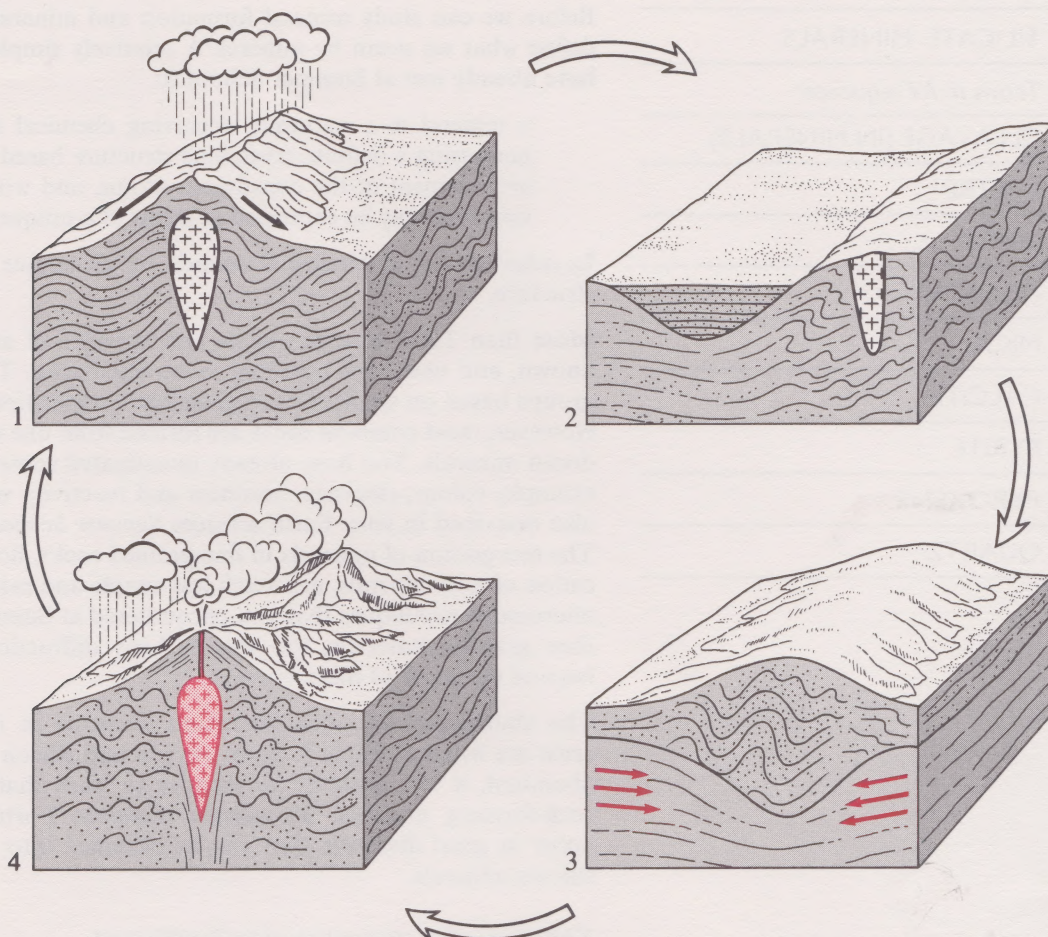


FIGURE 2 The rock cycle, as envisaged by Hutton: (1) existing rocks are eroded, (2) deposited to form sediments, (3) thrown up by earth movements into a mountain belt, (4) subjected to igneous and metamorphic processes, and then eroded once more.

2 ROCK MATERIALS

2.1 ROCKS AND MINERALS: AN INTRODUCTION TO THE AV SEQUENCE

Have you ever stopped to consider precisely what is meant by a **rock**? This word has probably been part of your vocabulary since childhood, but it is unlikely you have ever been asked to define it. You have already examined some of the rocks from your Kit, and others at Summer School. It was demonstrated in the AV sequence for Units 5–6 that rocks are composed of discrete grains that are either cemented together or crystallized into an interlocking mass. These grains are called **mineral** grains, and most rocks are simply consolidated aggregates of mineral grains.

Igneous rocks, which form by crystallization from a melt, actually contain several kinds of mineral, as you can see for yourself in Kit samples S1 and S5. Some sedimentary rocks, however, are composed predominantly of one mineral, reflecting some process of selection in the rock cycle that has favoured its inclusion. Sample S6, for example, is made up almost entirely of quartz grains, and sample S7 contains a high proportion of the mineral calcite. The assemblage of minerals found in a rock is not, therefore, random but directly related to the original rock-forming processes—processes that we shall be exploring in this Unit.

SILICATE MINERALS

Terms in AV sequence:

CLEAVAGE (IN MINERALS)

LUSTRE

SILICA

FELDSPAR

MICA

CALCITE

PYRITE

PYROXENE

QUARTZ

Before we can study mineral formation and mineral assemblages we must define what we mean by mineral. A relatively simple definition, which you have already met at Summer School is:

a mineral is a naturally occurring chemical element or compound, possessing a definite crystalline structure based on an ordered internal arrangement of the constituent atoms, and with a chemical composition that may be expressed in terms of a unique chemical formula.

In other words, a mineral is defined by its unique chemistry and atomic structure.

More than 2000 mineral species and varieties of mineral species are now known, and new ones continue to be discovered. These are classified into groups based on similarities in chemical composition and atomic structure. However, most common rocks are formed from one or more of fewer than a dozen minerals. You have already investigated some mineral properties, for example, colour, cleavage, hardness and reactivity with acids, and they are also described in your Earth sciences *Summer School Laboratory Notebook*. The recognition of minerals in fine-grained rocks, however, and the identification of rare minerals, are much less simple and may require the use of the microscope techniques that you encountered at Summer School or, for even finer grained materials, the use of X-ray diffraction methods (which are beyond the scope of this Course).

The abundances of some important elements in the Earth's continental crust are listed in Table 1. Since oxygen and silicon are so overwhelmingly abundant, it will come as no surprise to learn that all the most common rock-forming minerals are **silicate minerals**. Fortunately, although they occur in great diversity, they can be classified into a handful of groups of silicate minerals.

TABLE 1 The composition of the Earth's crust.

(a) Abundances by mass of the most common elements				(b) Abundances by mass of some elements used in everyday life		
		%	parts per million			parts per million
oxygen	O	46.6	466 000	sulphur	S	520
silicon	Si	27.7	277 000	carbon	C	320
aluminium	Al	8.3	83 000	chlorine	Cl	200
iron	Fe	5.0	50 000	nickel	Ni	80
magnesium	Mg	2.1	21 000	zinc	Zn	65
calcium	Ca	3.6	36 000	nitrogen	N	46
sodium	Na	2.8	28 000	copper	Cu	45
potassium	K	2.6	26 000	lead	Pb	15
TOTAL		<u>98.7</u>		tin	Sn	3
				uranium	U	2
				mercury	Hg	0.5
				silver	Ag	0.1
				gold	Au	0.005

You should now listen to the AV sequence 'Minerals' (Tape 4, Side 1, Band 2), which introduces some of the more important properties of the mineral samples in the Kit. For this sequence you will need from the Kit the rock and mineral samples, the hand lens, hydrochloric acid (and a tissue), a pipette and the watch glass. You will also need a copper coin and a pencil. During the sequence you will be referred to Plates 5 to 9, which are at the back of this binding. While working through the tape you should be ready to complete Table 2.

In the next Section we shall examine the structure and chemistry of the silicate minerals in more detail.

TABLE 2 Main physical and chemical properties of some common rock-forming and ore minerals. For use with the AV sequence 'Minerals'.

Specimen no.	Mineral	Composition	Colour	Lustre	Cleavage	Hardness	Acid test (HCl)
I		SiO_2	clear—can be grey, pink, purple	glassy	none (irregular fracture)	harder than steel	
II		KAlSi_3O_8 or $\text{CaAl}_2\text{Si}_2\text{O}_8$					
III		$\text{KAl}_3\text{Si}_3\text{O}_{10}(\text{OH})_2$			one perfect cleavage		no reaction
IV		CaCO_3					
V		FeS_2		metallic, shiny		harder than steel	
not supplied	olivine	$(\text{Mg, Fe})_2\text{SiO}_4$	pale green when fresh, to yellowy-brown	glassy	none	hard—similar to quartz	no reaction
not supplied	pyroxene	$(\text{Mg, Fe, Ca})_2\text{Si}_2\text{O}_6$	dark green to black	glassy	two at right angles	quite hard—similar to feldspar	no reaction

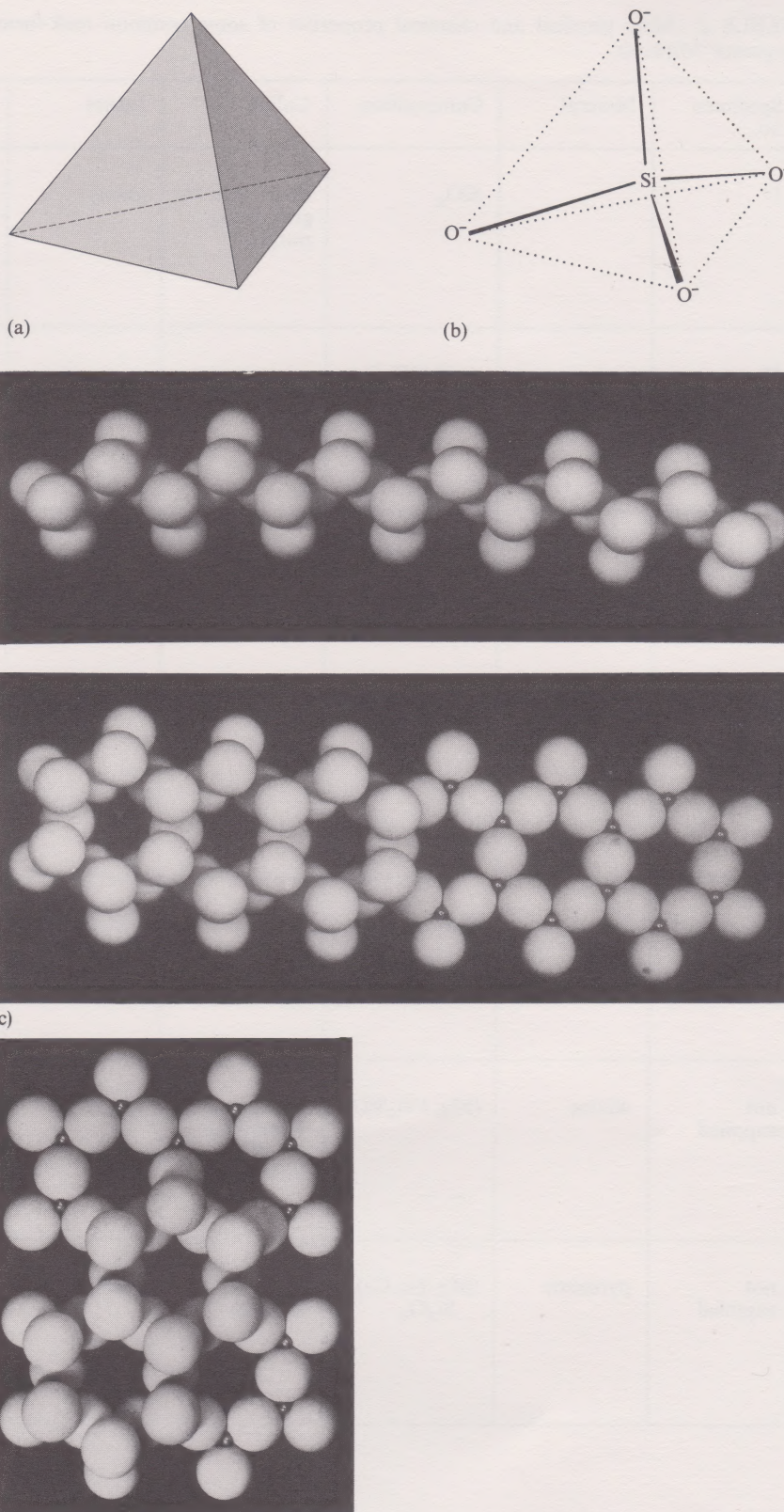
2.2 THE SILICATE MINERALS

The atomic structure of these minerals is based upon the silicate unit $(\text{SiO}_4)^{4-}$ (Figure 3), which can be represented as a sort of tetrahedral building block. (A tetrahedron is a four-sided body, like a pyramid.) The unit has four residual negative charges that are balanced either by appropriate cations, such as Fe^{2+} , Mg^{2+} , Na^+ , K^+ and Al^{3+} , or by the sharing of one or more oxygen atoms with other silicate units. The different ways in which the tetrahedral silicate building blocks are linked together, and the choice of cations to balance the negative charges, are responsible for the variety of silicate mineral species. The tetrahedra can be linked in chains, sheets or more complex frameworks (Figure 3c).

OLIVINE

FERROMAGNESIAN MINERALS

FIGURE 3 (a) A tetrahedron, the basic shape of the $(\text{SiO}_4)^{4-}$ unit. (b) The atomic structure of the $(\text{SiO}_4)^{4-}$ unit to show how the silicon and oxygen atoms are linked. (c) The structure of some other silicates: a chain (e.g. pyroxene), a double chain (e.g. amphibole) and a sheet (e.g. mica). The models are to scale, with large oxygen atoms and small silicon atoms.



Six major sub-groups of silicate minerals are recognized, based upon the way in which the silicate units are joined to each other. These groups and their chemical compositions are listed in Table 3. The elements that are grouped together in round brackets are alternatives that can occur in place of each other. For example, the **olivines** may contain *all* magnesium (Mg_2SiO_4) or *all* iron (Fe_2SiO_4), but more commonly they contain a mixture of the two in any proportion. Some of these minerals should now be familiar to you, since you examined them during the AV sequence 'Minerals', and the rest you will have seen in the Earth sciences laboratories at Summer School.

TABLE 3 The structure and chemical compositions of the most common sub-groups of silicate minerals in rocks.

	Sub-group	Most common colour in rocks	Chemical composition	Temperature of crystallization
Iron- and magnesium-rich minerals	OLIVINES	pale green (Mg rich) to yellowy-brown (Fe rich)	$(\text{Mg, Fe})_2\text{SiO}_4$ (the Mg-rich variety is the gemstone peridot)	↑ HIGHEST
	PYROXENES	dark green to black	$(\text{Ca, Mg, Fe})\text{Si}_2\text{O}_6$	
	AMPHIBOLES	dark green to black	a hydrated aluminium-silicate containing Ca, Na, Mg and Fe	
	MICAS	brown (Fe-rich) biotite to white (Fe-poor) muscovite	a hydrated aluminium-silicate containing K, Mg and Fe a hydrated aluminium-silicate containing K	
Iron- and magnesium-poor minerals	FELDSPARS	white or sometimes pink—plagioclase	$\text{CaAl}_2\text{Si}_2\text{O}_8$ and $\text{NaAlSi}_3\text{O}_8$ (or any mixture of Ca and Na)	↓ LOWEST
	QUARTZ	white or pink—potassium feldspar colourless	KAlSi_3O_8 SiO_2	

Large cations, such as K^+ , Ca^{2+} and Na^+ , found in amphiboles, micas and feldspar are associated with more 'open' atomic structures. In some cases, such as the amphiboles and micas, there are some gaps in the structure that are just the right size for hydroxyl ions, OH^- , to be accommodated. These minerals are, therefore, said to be *hydrated*.

- ☐ Do you notice in Table 3 a difference between the colours of minerals that are rich in iron and magnesium (i.e. **ferromagnesian minerals**), and those that are poor in these elements?
- ☒ The minerals that are rich in iron and magnesium are all green, brown or black, whereas the others are all pale coloured, white or colourless, because it is the presence of iron that makes minerals dark.

2.3 THE CRYSTALLIZATION OF SILICATE MINERALS

The approximate chemical compositions of some common crustal rocks, similar to those in your Kit, are listed in Table 4. Don't be overawed by all the numbers in this Table, you will not have to recall them. Note that the Table is expressed in terms of element *oxides*. This is conventional, for historical reasons, when tabulating the composition of minerals and rocks. It is most important that you should *not* conclude that elements are actually present as oxides in all cases. Some oxides such as TiO_2 may indeed be present in that form, but others, such as K_2O would be chemically unstable, and the potassium is always bound into more stable compounds, usually silicate minerals.

TABLE 4 Typical chemical composition by mass (%) of some common crustal rocks that are comparable to those in the Kit.

	Peridotite	Basalt	Andesite	Granite	Sandstone
SiO_2	46	49	56	72	81
TiO_2	0.13	1.4	0.7	0.4	0.1
Al_2O_3	2.7	16	16	15	7.6
$\text{Fe}_2\text{O}_3 + \text{FeO}$	7.3	9.4	7.8	3.4	4.3
MgO	42	8.5	6.4	0.9	—
CaO	2.0	11	7.9	2.0	—
Na_2O	0.22	2.7	2.9	3.5	0.6
K_2O	0.03	0.26	1.6	4.2	4.8
TOTALS*	100	98	99	101	98

* The totals of the analyses are not 100% because of rounding errors and various uncertainties.

PERIDOTITE

PLAGIOCLASE

BASALT

You now have some chemical data on rocks, and on the minerals that make up rocks. Each rock is made up of one or more of the different kinds of silicate mineral. By inspecting the two sets of data in Tables 3 and 4, you might get the impression that you could arrive at *any* of the rock compositions given, simply by mixing different minerals in the right proportions. Thus, by juggling numbers, it might seem possible to arrive at a granitic composition by mixing quartz, pyroxene and feldspars. The fact is, however, that this simple mixing does *not* take place, and each type of rock is always characterized by a distinctive mineral composition. **Peridotites**, for example, invariably contain olivine and pyroxene, whereas olivine is *never* found in granites.

The main physical factor that determines whether a particular mineral is present after cooling of molten material, or *magma*, is the *crystallization temperature*; but there's also an important constraint dictated by the rock's chemical composition. Each mineral is characterized by a different crystallization temperature; hence, when a magma begins to cool, the silicate minerals will begin to crystallize in a sequence that is controlled by these temperatures. The approximate order in which silicate minerals crystallize is shown in the right-hand column of Table 3, and also in Figure 4; but both the Table and the Figure represent a considerable simplification of the complex natural situation. The order of crystallization can be affected by other physical factors, such as pressure, and by the chemical composition of the magma. Furthermore, the sequence is a generalized one; it is *not* intended to show which minerals will crystallize when any specific rock melt is cooled. The laws of chemistry dictate that only a small selection of silicates can nucleate in a magma of any specific composition.

From Figure 4 you can see that the Ca-bearing **plagioclase** feldspars are able to crystallize at the same temperature as some of the ferromagnesian minerals. In contrast, mica, alkali (Na and K) feldspars and quartz are generally among the minerals that crystallize at the lowest temperatures.

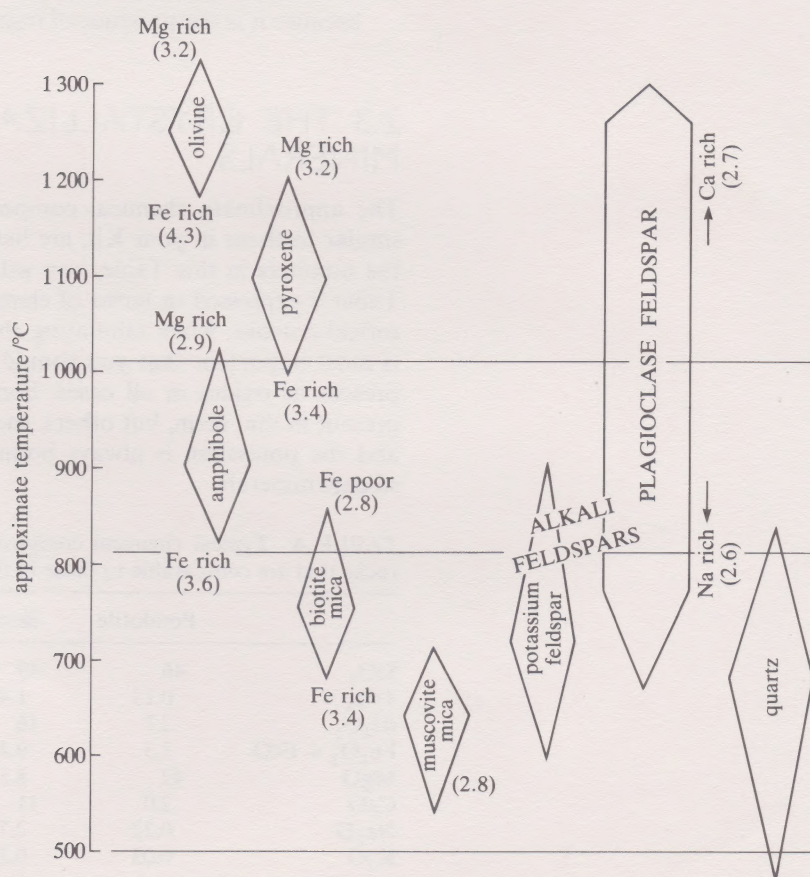


FIGURE 4 Ranges of crystallization temperatures for common minerals of igneous rocks. The figures in brackets are densities ($\times 10^3 \text{ kg m}^{-3}$).

You should now be able to see how many of the properties of the silicate minerals are related to each other, and how they change with decreasing temperature of crystallization. For example, you have already seen that the *colour* of a mineral is affected by its *chemical composition*, and that chemical composition is controlled by the *atomic structure* of the mineral. You should now be able to work out how *chemical composition* and *atomic structure* are also related to the *temperatures of crystallization* of a mineral.

ITQ 1 (a) How do the proportions of magnesium and iron change with temperature *within* any individual mineral group, such as the olivines?
 (b) How do the proportions of calcium and sodium change with temperature *within* the plagioclase-feldspar group?
 (c) How do the Fe, Mg, Si, Na and K contents of silicate minerals change *in general* with falling temperature of crystallization?

Remember that groups of minerals within a particular temperature band (Figure 4) crystallize together to form igneous rocks. So, for example, **basalts** comprise pyroxene and feldspars whereas granites are made up of quartz, feldspars and micas. Just how this relates to the formation and composition of the parent magmas is the subject of Section 3. As you will see in Section 4.2, another important consequence of the order of crystallization temperatures is that it influences the way in which the silicate minerals in igneous rocks break down to form raw materials for the sedimentary rocks.

2.4 THE DENSITIES OF SILICATE MINERALS

We can now begin to make some deductions about the *densities* of minerals and, consequently, about the densities of different types of rock. The approximate relative atomic masses of the most common elements in silicate minerals are listed in decreasing order in Table 5. In what follows, we shall make the simplifying assumption that the atoms of different elements are all the same size, so that we can use relative atomic mass as a guide to the density of an element.

If you compare the relative atomic mass of iron with that of magnesium, it is quite easy to see why the iron-rich members of any mineral group are denser than the magnesium-rich members of the same group. Similarly, if you compare the relative atomic mass of calcium with that of sodium, it is clear why calcium-rich plagioclase feldspars are more dense than sodium-rich varieties. It is much more difficult to interpret differences in densities *between* mineral groups.

- ☐ What factor, other than the relative atomic mass of its constituent atoms, should affect the density of any crystalline material?
- ☒ Clearly, the more closely packed the atoms are in a structure, the denser that material will be.

A good example of the effects of atomic packing is provided by graphite and diamond (Units 13–14, Section 7). Both are made of pure carbon, but diamond has a much higher density due to the close packing of its structure. Rocks that are made up of high-density minerals naturally have high densities themselves; but how can we measure the density of the mineral or rock samples in the Kit?

You may remember from Unit 3 (Section 6.2) that

$$\text{density} = \frac{\text{mass}}{\text{volume}}$$

The samples in the Kit are small, so it is more convenient to work in units of g and cm³, rather than in the SI units of kg and m³.

TABLE 5 The approximate relative atomic masses of the common elements of silicate minerals.

Element	Relative atomic mass
iron (Fe)	56
calcium (Ca)	40
potassium (K)	39
silicon (Si)	28
aluminium (Al)	27
magnesium (Mg)	24
sodium (Na)	23
oxygen (O)	16

ITQ 2 The density of water is such that one cubic metre has a mass of one tonne (1000 kg), i.e. its density is $1 \times 10^3 \text{ kg m}^{-3}$. What is its density in g cm^{-3} ?

The mass of each sample is easy to measure accurately, using the balance. Volume is not as easy, because the rock and mineral samples are irregular in shape.

- ☐ If you were to put a cube of rock, volume 10 cm^3 , into a graduated beaker containing 50 cm^3 of water, what would happen to the level of the water?
- ☒ The water level would rise to the 60 cm^3 mark on the graduated scale of the beaker, because the volume of the cube would displace an equal volume of water.

But even using a calibrated vessel, direct measurement of the volume cannot be done very accurately; we shall use a method based upon mass, which we can measure accurately with the balance.

ITQ 3 (a) If an object displaces 10 cm^3 of water and the density of water is 1 g cm^{-3} , what is the *mass* of the displaced water?
 (b) Assuming that the acceleration due to gravity $g = 1000 \text{ cm s}^{-2}$, what is the *weight* of the displaced water?

You will be aware that the weight of an object *appears* to be less when it is immersed in water, because of the upwards supporting force of the water. *Archimedes' principle* tells us that the magnitude of the upwards supporting force is equal to the magnitude of the weight of the displaced water (Units 7–8). So the difference in the weight of a rock in air and its weight in water must be equal to the weight of the displaced water. Expressed mathematically:

$$\text{weight of displaced water} = \text{weight of rock in air} - \text{weight of rock in water}$$

Converting to mass (since acceleration due to gravity is constant, weight is proportional to mass):

$$\text{mass of displaced water} = \text{mass of rock in air} - \text{apparent mass of rock in water}$$

We can use the balance to determine the mass of any rock sample and the mass of the water it displaces. Since the density of water is 1 g cm^{-3} , the volume (in cm^3) is numerically the same as the mass (in g). So the decrease in measured mass due to the immersion of the sample in water gives us a direct measurement of the magnitude of its volume:

$$\begin{aligned} \text{volume of sample} &= \text{volume of displaced water} \\ &= \frac{\text{mass of displaced water}}{\text{density of water}} \\ &= \frac{\text{mass of rock in air (g)} - \text{apparent mass of rock in water (g)}}{1 \text{ g cm}^{-3}} \quad (1) \end{aligned}$$

or, magnitude of volume of sample (in cm^3) = magnitude of apparent mass loss (in g)

So, by weighing a sample in air and in water we can obtain all the quantities needed to calculate the density of a sample; this is the basis of the investigation, which you should now complete.

EXPERIMENT

TIME

You should be able to complete this experiment in $1\frac{1}{4}$ – $1\frac{1}{2}$ hours. (Note: you should check the assignment booklet before starting the experiment as assignment questions may involve this method. You will save yourself time if you examine all the required samples in one session.)

KIT ITEMS

Part 1
granite sample and any other rock samples specified in the assignment booklet

Part 2
balance
100 cm³ beaker, three-quarters filled with water
quartz and pyrite samples and any other mineral samples specified in the assignment booklet

NON-KIT ITEMS

screwdriver to adjust the balance
tissues
cup if the samples will not fit the 100 cm³ beaker
calculator
something with which to hang your specimens from the balance beam: fuse wire or sewing cotton give the most accurate result; a piece of plastic netting (e.g. from oranges, onions or garlic) or a hairnet are easier but less accurate because there is more extraneous material to be weighed and immersed

INVESTIGATING THE DENSITY OF ROCKS AND MINERALS

METHOD

Set up the balance, remind yourself how it is used and zero it; have the beaker of water to hand.

Carry out the following procedure, first for the quartz and pyrite.

- 1 Measure the mass of the sample by placing it on the pan of the balance.
- 2 Record the mass of the sample in Table 6a.
- 3 Suspend the sample from the beam of the balance, using wire, cotton or netting. The sample should hang about 1 cm above the pan, which should not be removed (Figure 5a).
- 4 Measure the mass of the sample plus wire, cotton or netting (Figure 5a) and record it in Table 6a. It should be a little more than the mass of the sample alone (say up to 1 g), corresponding to the mass of the wire, cotton or netting.
- 5 Take the 100 cm³ beaker with water in one hand and place the sample in the water being careful
(a) not to let the beaker touch the pan or the balance support, and
(b) not to let the sample touch the beaker sides.
- 6 Holding the beaker with one hand, measure the mass of the sample while it is totally immersed (Figure 5b). It's a bit tricky but you'll soon get the hang of it; if you find the beaker difficult to support manually, you could try constructing a 'bridge' over the scale-pan to support the beaker (Figure 5c). Air bubbles trapped around the specimen increase its buoyancy and thus the apparent mass loss will be greater. If you can see any air bubbles, shake the sample slightly to dislodge them. Similarly, remove any drops of water on the scale-pan with a tissue, as these will increase the apparent mass of the sample. What is the apparent mass now?
- 7 Record the immersed mass in Table 6a.

TABLE 6a The density of mineral samples.

Sample	quartz	pyrite	olivine	feldspar
1 Mass of sample in air/g			82.31	7.68
2 Mass of sample + wire in air/g			82.34	7.71
3 Mass of sample + wire in water/g			57.70	4.74
4 Volume/cm ³			24.64	
5 Density/g cm ⁻³			3.34	
6 Density/10 ³ kg m ⁻³			3.34	

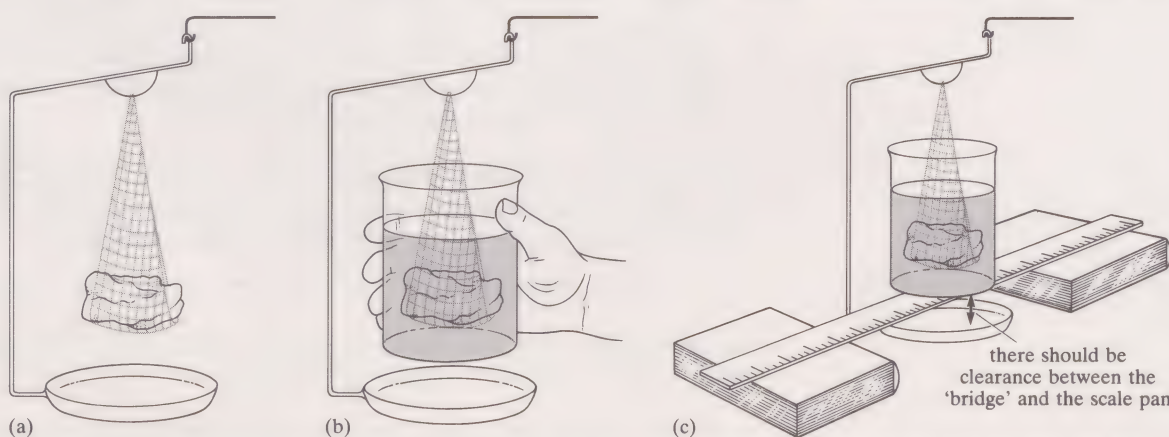


FIGURE 5 (a) Weighing the sample and wire (or netting). (b) Weighing the sample in water, making sure the sample does not touch the sides or bottom of the beaker. (c) Making a 'bridge' to support the beaker, with two books and a ruler.

EXPERIMENT
CONTINUED

The magnitude of the volume of the sample (in cm^3) is equal to the magnitude of the apparent loss of mass (in g); that is, in Table 6a, the value in row 4 equals that in row 2 minus that in row 3, and the units will be cm^3 . The density of the sample, in g cm^{-3} , is the original mass (row 1) divided by the volume (row 4); to convert to density in kg m^{-3} , multiply by 10^3 .

You will see in Table 6a that we have done the measurements and calculations for one sample, olivine, and the measurements for another, feldspar, so that the arithmetic will be clear to you. Repeat the procedure in steps 1 to 7 and complete Table 6a. You might prefer to carry out all the measurements before doing the calculations for the specimens.

- ITQ 4** (a) Calculate the density of feldspar from the data in Table 6a.
 (b) Reviewing progress so far, what values have you obtained for the density of quartz and pyrite?
 (c) How do these compare with the values for feldspar and olivine?

Now try a rock sample.

Using the same experimental procedure and completing Table 6b, find the density of the granite. If the sample is too large for the beaker, improvise by using a cup, or something similar.

TABLE 6b The density of rock samples.

Sample	granite	peridotite	gabbro
1 Mass of sample in air/g		64.41	115.14
2 Mass of sample + wire in air/g		64.70	115.50
3 Mass of sample + wire in water/g		45.10	77.34
4 Volume/ cm^3		19.60	38.16
5 Density/ g cm^{-3}		3.29	3.02
6 Density/ 10^3 kg m^{-3}		3.29	3.02

Once you get the hang of the procedure you might like to try other substances: common iron nails work well and should give you a density of about $7.87 \times 10^3 \text{ kg m}^{-3}$. Or, for an even higher density, you could try a piece of gold or silver, for example, a ring.

- ☐ Why does a 'gold' object, such as a ring, not have the density of pure gold ($19 \times 10^3 \text{ kg m}^{-3}$)?
- ☒ Because 'gold' jewellery is usually only 30–75% pure (called 9, 14 or 18 carat; pure gold is 24 carat). The added material, for example, copper or silver, is less dense and reduces the overall density. We obtained a density of $13.8 \times 10^3 \text{ kg m}^{-3}$ for a gold ring, using the method above. Incidentally, this is one of the simplest ways of measuring the purity of a gold sample and was the reason Archimedes was trying to measure density in the first place!

You have now completed the practical part of this experiment; if there are assignment questions involving other specimens you should measure the densities of these now, before putting away the equipment.

Let us consider what the densities you have obtained in Table 6a and 6b imply for the composition of parts of the Earth.

- ☐ Why, in terms of mineral densities, is the density of granite less than that of peridotite?

- Because granite is composed mostly of feldspar and quartz, which have densities that are far lower than the ferromagnesian minerals, such as olivine and pyroxene, that make up peridotite.
- If two minerals of different densities, say olivine ($3.3 \times 10^3 \text{ kg m}^{-3}$) and calcium-rich plagioclase ($2.7 \times 10^3 \text{ kg m}^{-3}$), are crystallizing together from a melt of density $2.7 \times 10^3 \text{ kg m}^{-3}$, what will happen to the olivine and plagioclase crystals formed?
- The olivine crystals are denser than the melt, so they will tend to sink in the magma chamber, but the plagioclase crystals, being about the same density as the melt, will remain approximately where they formed.

In fact, the situation is a bit more complicated than this because the composition and density of the melt changes during crystallization, and we explain this in more detail in Section 3.4. You will find that your basic knowledge of the density of minerals is useful at several points in this Unit, particularly when we discuss the formation of sedimentary and igneous rocks. The next Section examines the distribution of silicate minerals in the three principal types of crustal rock.

2.5 THE DISTRIBUTION OF SILICATE MINERALS IN CRUSTAL ROCKS

So far, we have considered the major groups of silicate minerals that occur in igneous rocks. We will now take a preliminary look at the mineralogy of sedimentary and metamorphic rocks. You already know something about sedimentary rocks from the first AV sequence in Units 5–6 but, so far, little has been said about metamorphic rocks. They are simply rocks that have experienced changes. Initially they may have been either igneous or sedimentary rocks; subsequently they were subjected to heat and pressure to such an extent that they became recognizably different from their parent rocks.

Table 7 lists the main groups of silicate minerals found in igneous, sedimentary and metamorphic rocks. Two of these mineral groups, clay minerals and garnets, have not been discussed before. Clay minerals, as their name suggests, are the chief constituent of clays. They are hydrated aluminium silicates, somewhat similar to micas in their structure but of lower density and generally with finer mineral particles. Garnets are familiar as large red gemstone crystals and as industrial abrasives. They are usually ferromagnesian aluminosilicates with rather dense atomic packing, similar to that in the olivines; therefore they are dense minerals.

TABLE 7 The silicate mineral groups found in igneous, sedimentary and metamorphic rocks.

Igneous rocks	Sedimentary rocks	Metamorphic rocks
olivines	clay minerals	garnets
pyroxenes		
amphiboles		amphiboles
micas	micas	micas
feldspars	feldspars	feldspars
quartz	quartz	quartz

ITQ 5 From Table 7 identify the principal differences between the minerals of the three rock groups. Are the minerals of igneous rocks more like those of metamorphic rocks than those of sedimentary rocks?

IGNEOUS ROCKS

The reasons for the differences in mineral assemblages between the major rock types are fundamental, and are due to the operation of different processes within the rock cycle. On exposure at the Earth's surface, many rocks are out of equilibrium with the conditions under which they formed. Igneous or metamorphic rocks are subject to much higher temperatures and pressures during formation than those occurring at the Earth's surface. Many minerals in these rocks are, therefore, unstable at the Earth's surface. Some minerals are mechanically broken down into fragments, and others are chemically altered by surface geological processes. The products of both processes are involved in the formation of sedimentary rocks, which takes place at the surface of the Earth, at low temperatures. When sedimentary rocks are metamorphosed, they are heated up; pre-existing minerals may be destroyed, and new ones formed. Since the physical conditions of metamorphism are much more like those involved in the formation of igneous rocks than of sedimentary ones, it is only to be expected that metamorphic minerals show similarities to igneous ones. You will be studying the parts in the rock cycle that lead to the formation of the major rock groups in Sections 3, 4 and 5.

SUMMARY OF SECTION 2

- 1 Rocks are consolidated or crystalline aggregates of mineral grains. There are many varieties of mineral species known, but fewer than a dozen of these are the main rock-forming minerals. Most of these are silicates.
- 2 The main building block of silicates is the $(\text{SiO}_4)^{4-}$ tetrahedron, and the linking of tetrahedra at the molecular level results in different internal mineral atomic structures; micas, for example, have a sheet-like structure.
- 3 Minerals that are rich in iron and magnesium (ferromagnesian minerals), such as olivine, pyroxene, amphibole, garnet and brown mica, tend to be dark in colour (black, dark green, brown or dark red). Minerals that are poor in iron and magnesium (e.g. quartz, feldspar, white mica) tend to be pale in colour.
- 4 Crystallization from silicate melts results in a sequence of minerals, each characterized by a different crystallization temperature. Minerals from a particular temperature band tend to occur together in igneous rocks. The variety of these rocks, therefore, reflects differences in crystallization temperature as well as differences in chemical composition.
- 5 Ferromagnesian minerals are denser than those that are poor in iron and magnesium. Within one group of minerals, iron-rich forms are denser than magnesium-rich ones because iron atoms are denser than magnesium atoms.
- 6 The density of minerals, rocks and other solids can be determined experimentally, and this may give an indication of the chemical composition of a sample.

SAQ 1 Select from your Kit the coarse-grained igneous rock samples: S1 (granite), S4 (peridotite) and S5 (gabbro). Make sure that you have a strong light to shine on the samples, and a hand lens or magnifying glass available to see the individual minerals more clearly.

- (a) These three rocks and their average mineral compositions are listed in Table 8. Using your rock samples, fill in the empty spaces in Table 8. (Remember that the details will vary from sample to sample, and that the mineral compositions quoted are averages for the rock type. You should complete the Table *from your own observations*. Note whether the details differ slightly from those in the completed Table in the answer to this SAQ, but don't worry about minor discrepancies.)
- (b) Why is there an overall colour difference between rock sample S1 and rock samples S4 and S5?
- (c) What are the reasons for the density differences between sample S1 and samples S4 and S5?

TABLE 8 For use with SAQ1.

Rock sample	Average mineral composition	Colour of mineral	Overall rock colour (light or dark)	Density/ kg m^{-3}	Relative temperature of formation (high, moderate, low)	Expected sequence of mineral crystallization
S1 GRANITE	10% biotite mica					
	60%, mainly K-feldspar					
	30% quartz					
S5 GABBRO	10% olivine					
	30% pyroxene					
	60% plagioclase feldspar					
S4 PERIDOTITE	85% olivine					
	15% pyroxene					

(d) The feldspar crystals in sample S1 have good, rectangular crystal outlines whereas the quartz crystals are irregular. Why might this be so?

(e) What can you say about the *rate* of cooling and crystallization of all three samples?

SAQ 2 Take the basalt sample, S3, from your Kit.

(a) From the overall *colour* of the rock, which types of mineral would you expect to be present?

(b) Can you suggest whether the *relative* temperature of formation of S3 was high, moderate or low, on the scale of Figure 4?

(c) What can you say about its rate of cooling and crystallization?

(d) You have measured the density of sample S1 (granite) and you were given the values for S4 (peridotite) and S5 (gabbro). Which of the samples in Table 8 do you think S3 will resemble most in density?

Note: It is important that you read the answers to SAQ 1 and SAQ 2 before continuing to the next Section.

3 IGNEOUS PROCESSES

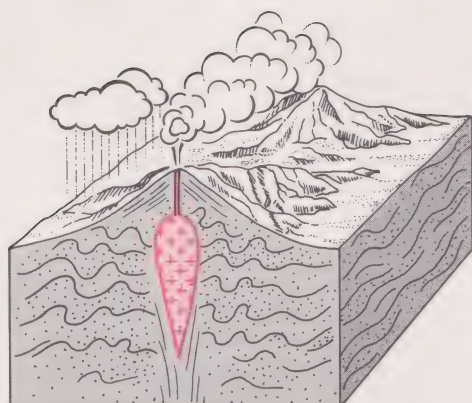
Igneous rocks form the starting point in the rock cycle: the materials making up all sedimentary and metamorphic rocks were derived originally from an igneous source, although for any given rock this primitive source may be lost in the mists of time. In this Section, we shall be examining the essential part that igneous rocks play in the rock cycle, but we shall also investigate how igneous rocks themselves are formed, and try to answer some important questions, such as:

Why is there such a wide range of igneous rock types?

What processes are involved in generating one rock from another?

Why are quite different kinds of lava erupted from volcanoes at constructive plate margins compared with those erupted at destructive plate margins?

The ultimate source of all **igneous rocks** lies in the mantle, so it is there that we shall start.



PYROCLASTIC ROCKS

3.1 THE MANTLE

70% of the Earth's mass and 80% of its volume consists of mantle rocks. Arguably, therefore, we ought to devote most of this Unit to the mantle. Unfortunately, it is inaccessible and consequently we have to rely on indirect evidence, such as seismic techniques and inferences drawn from surface observations of rocks and minerals.

- ☐ Which of your rock samples is similar to mantle rock?
☒ The peridotite, S4, is similar to mantle rock.

The chemical composition of the peridotite samples found at the surface of the Earth has been exhaustively studied; as you have already learned, peridotite is composed mainly of the mineral olivine, a ferromagnesian silicate. You may also recall from Units 7–8 that the process of sea-floor spreading is connected with convection currents in the mantle, which rise up and spread outwards beneath the ocean ridges.

Let's think about what the source of heat energy that 'drives' these convection cells might be. The most likely source of the Earth's internal energy is decay of long-lived radioactive isotopes of the elements potassium (K), uranium (U) and thorium (Th); of these, potassium is by far the most common. Since the mantle is thought to be in a state of constant convective motion, you might suppose that it is rich in these elements with radioactive isotopes. Look back to Table 4 (page 9) and answer the following questions.

- ITQ 6** (a) Which two of the rocks contain the most potassium and which contains the least?
 (b) What is the ratio of potassium in the richest rock to that in the poorest rock?
 (c) Are the rocks that are richest in potassium continental crustal, oceanic crustal or mantle rocks?

As you have seen, peridotite is dominantly Mg_2SiO_4 and contains only a minute amount of potassium. In fact, there is sufficient of the radioactive potassium isotope to provide the heat energy that drives convective cells in the mantle. Now let's consider the rarer elements, uranium and thorium, as well. Table 9 gives the abundance of the radioactive elements uranium, thorium and potassium in granite, basalt and peridotite, and the amount of heat produced from them.

Table 9 shows just what a poor producer of heat peridotite is, especially when compared with granite. Calculations show that the Earth releases 3×10^{13} watts continuously and the mantle produces 1.3×10^{13} watts, so the mantle contributes just over 40% of the total heat flow.

TABLE 9

Rock	Abundance of radioactive element/p.p.m.*			Heat produced/ $\text{J kg}^{-1} \text{s}^{-1}$
	uranium	thorium	potassium	
granite	4	20	40 000	10.3×10^{-10}
basalt	0.5	2	15 000	1.5×10^{-10}
peridotite	0.02	0.06	200	4.1×10^{-12}

* Parts per million. Because of the small amounts of the elements involved, they are expressed as the ratio of the element to one million parts of rock.

Clearly, crustal rocks, particularly granites, contribute a much greater amount of heat energy per unit mass. However, the mass of the crust is much less than that of the mantle and it is estimated that the total heat produced in the mantle *and* the crust accounts for about 50–75% of the

heat flow observed at the Earth's surface today. This shortfall has encouraged speculation that some of the Earth's interior heat energy, which drives mantle convection currents, has a different source.

One possibility is that the Earth retains a part of its *original* internal energy, known as primordial heat, which was trapped inside when the Earth formed. Rocks are such extremely poor conductors of heat that not all of this internal energy produced during the accretion of the Earth over four billion years ago would have leaked away into space. Another possibility is that the Earth's core may contain a tiny but significant amount of potassium, and that decay of the radioactive isotope of potassium in the core would contribute the extra energy necessary to account for present-day heat flow. So, on balance, a small contribution from primordial heat, in addition to the heat being generated today, is probably necessary to account for the Earth's observed heat flow.

3.2 VOLCANOES AND VOLCANIC ROCKS

One of the most obvious consequences of the Earth's hot, mobile interior is volcanic activity at the surface. You will already have seen something of volcanic rocks in the AV sequences and TV programmes of Units 5–6 and 7–8. Volcanic rocks scarcely need any definition—they are the products of volcanoes. But they have three important characteristics:

- 1 they crystallize at the surface of the Earth;
- 2 they are usually fine grained;
- 3 they do *not* cut across older rocks, but rest on top of them.

Volcanic rocks come in a wide range of shapes and forms. Your Kit specimens S2 and S3 are both volcanic rocks—basalts; but one (S3) was erupted as liquid lava and the other (S2) as a vesiculated (bubbly), low-density fragment that was probably hurled into the air in a volcanic eruption and subsequently fell to Earth to form part of a cindery, ash-like deposit (Plates 10a and b). Such fragmented igneous rocks are known as **pyroclastic** (fire-broken) **rocks**. Despite their marked *physical* differences, however, S2 and S3 are almost identical in *chemical* composition. Volcanic rocks of many other compositions are known, and these may occur either as lavas or as pyroclastic rocks.

When basalt magma is first erupted, it is at a temperature of about 1 100 °C. At this temperature, its viscosity is fairly low, so it can flow in a liquid fashion. As it cools, it becomes stiffer and more viscous. In this respect, basalt magmas are similar to almost all other liquids, which get progressively more viscous as they cool. You will certainly have noticed how sluggish a car engine is on a crisp, winter morning, when the engine oil is thick and viscous.

When a high-temperature, low-viscosity basalt magma approaches the surface, the gases dissolved within it (e.g. steam, CO₂ and SO₂) come out of solution like the bubbles in a beer bottle when the top is taken off. Because the magma is of low viscosity, the gas can blow off freely. The escaping gas often sprays showers of volatile liquid basalt into the air as *fire fountains*. Ultimately this forms basaltic material like S2. Such fountains are usually directly associated with free-flowing lava flows erupted laterally from a vent. So, generally speaking, basaltic eruptions are not violently explosive, and there is more lava erupted than ash.

The low viscosity of basaltic magma controls the shape of the volcanic structures that it builds up. Basaltic lavas spread out widely around the vents from which they are erupted. Because they spread out so widely, basaltic lavas accumulate to form broad, gently sloping shield volcanoes. The best-known examples of these are in the Hawaiian islands and Iceland, as you saw in TV8. Although the slopes on the Hawaiian islands rarely exceed a few degrees, they are the biggest volcanoes in the world and, taking account of the part under water, they can also be considered the

COMPOSITE CONES

PARTIAL MELTING

highest *mountains* in the world, since they rise over nine kilometres from the sea floor. (Everest and K2 are only about 8.8 kilometres above sea level.)

Not many basaltic volcanoes have the gentle slopes that are characteristic of those in Hawaii. Many have the simple conical shape that is more commonly associated with volcanoes. These are **composite cones**, constructed from both lavas and pyroclastic rocks. They owe their elegant, sweeping profiles mostly to the fact that as the volcano is built upwards by material added through the vent at its summit, so erosion and the angle of repose of pyroclastic material on its flanks removes material and spreads it out radially (Figure 6 and Plate 11).



(a)



(b)

FIGURE 6 (a) Mount Fuji, Japan. This cone, of andesitic composition, towers about 3 500 m above the surrounding plain. Any large-scale eruption would be catastrophic for the towns and villages around the volcano. (b) A line of volcanic cones erupting basaltic ash and volcanic gases (mainly steam) in the Kamchatka area (north-west Pacific), seen here during early 1976. Note the loose volcanic ash banked up against the volcanologists' tent about two kilometres away from the erupting vent.

3.3 CONSTRUCTIVE PLATE MARGINS AND THE ORIGIN OF BASALTS

In Units 7–8 and TV 8 you saw that basalts erupt in enormous volumes from volcanoes along constructive plate margins. You also saw that the crust is very thin beneath the oceans—the mantle is only a few kilometres beneath the surface. It follows, therefore, that the basalts that have erupted from constructive plate margins have more than a little to do with the peridotites of the mantle. Table 4 (page 9) gives chemical analyses of typical ocean basalt and a peridotite; look carefully at these data.

- ☐ What are the main chemical differences between the two rocks?
- ☒ There is *much* more titanium, aluminium, calcium, sodium and potassium in the basalt than in the peridotite, and much less magnesium. The silica content of the basalt is also somewhat greater than that of the peridotite.

Now, in Units 7–8 it was stated that the ocean ridges are sited on top of ascending convection currents of partially molten mantle material.

- ☐ What composition would the molten liquid have if the peridotite of the mantle were *completely* melted?
- ☒ Obviously, it would have the *same* composition as the original peridotite.

If only *part* of the peridotite is melted, however, the results are quite different. If you melt a pure substance, the first drops of liquid are bound to be the pure substance; for example, pure ice melts to pure water. But rocks are not pure substances—they are composed of mixtures of minerals. When a peridotite is melted, the *first* drops of liquid are *not* simply molten peridotite.

A 'fresh' mantle peridotite consists mostly of the mineral olivine, with smaller amounts of pyroxene, a little plagioclase feldspar and minor amounts of other minerals.

ITQ 7 Bearing in mind what you have read in Section 2.3, can you suggest what the composition of the first drops of liquid will be when peridotite is melted?

So, the composition of the first liquid does not bear any *simple* relation to that of the solid rock. In the case of peridotite, if only about 10–20% of the material becomes liquid, its composition best approximates to a mixture of pyroxene and plagioclase feldspar with a small amount of olivine. The fact that **partial melting** yields liquids of different composition from the solid starting material is an extremely important one in geology, and it helps to explain the wide diversity of igneous rocks that occur in the Earth's crust.

The significance of this discussion of the partial melting of peridotite is that it is generally thought that basalts—which consist mostly of a mixture of pyroxenes, plagioclase feldspar and often a little olivine—are created at constructive plate margins by *partial melting of mantle peridotite*. The details of the melting process are complex and need not concern us here. The important fact is that the partial melting of mantle peridotite to form basaltic magmas provides all the material required to generate both the coarse-grained gabbros at depth and the extrusive basalts of the oceanic crust as sea-floor spreading proceeds.

- ☐ Look again at the analysis of basalt and peridotite in Table 4. By how much have the elements Ti, Na and K become concentrated as peridotite has yielded a basaltic magma?
- ☒ These elements are roughly ten times more abundant in basalt than in peridotite.

Does this suggest anything to you about what proportion of the peridotite has to be melted in order to make basalt?

The answer to this question may not be immediately obvious. But think about it this way. If all the K_2O in the peridotite (0.03%) goes into the basalt (where K_2O is 0.26%) the concentration factor is $0.26/0.03 = 8.6$. In other words, the K_2O that was originally in the peridotite has been distributed into a mass of basalt that is $1/8.6$ of the mass of the original peridotite. This is approximately 11%.

There are many other factors that have not been taken into account here, but calculations like these enable geologists to estimate the fraction of mantle peridotite that is melted in making basalts; in general, the fraction is less than 20%.

PLUTONIC ROCKS

PLUTON

CHILLED MARGIN

FRACTIONAL
CRYSTALLIZATION

CUMULATES

3.4 PLUTONIC ROCKS: FRACTIONAL
CRYSTALLIZATION AND PARTIAL MELTING

The basalts that are erupted by volcanoes at constructive plate margins are only the visible surface manifestations of processes taking place underground. **Plutonic rocks** are an important group of igneous rocks that originate as liquid magmas, like basalts, but are intruded or squeezed into other, older rocks. Plutonic magmas never reach the surface. Pluto was the mythological god of the Underworld, hence his name is well applied to rocks that only see the light of day long after they have crystallized. Plutonic rocks cool from their liquid state extremely slowly, since they are well insulated by surrounding rocks. As explained in the second AV sequence of Units 5–6, this means that when they crystallize and solidify, their crystals have plenty of time to grow.

It is clear that the centre of a large **pluton** (a large mass of plutonic igneous rock) will remain hot for an extremely long time—perhaps thousands of years, or maybe even a million years for a large granite—and, therefore, its centre will be very coarse grained. But what will happen at the *edges* of the pluton?

Clearly, where the pluton comes up against the surrounding rocks, it will be rapidly chilled, and a much finer grained rock will result on crystallization. The degree of chilling, of course, depends on how cool the surrounding rocks are; at great depths in the Earth's crust, there may be very little temperature difference between the intrusive magma and its host. Because rocks are such poor conductors of heat, the thickness of the fine-grained **chilled margin** is never very great; it may be a centimetre or less in a metre-thick dyke, and a metre or so in a pluton with a diameter of one kilometre.

There are thus three characteristics of plutonic rocks:

- 1 they cut across older rocks;
- 2 they are coarse grained, often with chilled margins;
- 3 they are formed underground.

So far, we have discussed rocks of basaltic composition only. But now consider what would happen if a large amount of basaltic magma, produced by partial melting of mantle peridotite, were intruded into a shallow part of the crust, say a couple of kilometres below the surface. The first thing that would happen, of course, is that a chilled margin would be produced all round the pluton, at the surface of contact with the surrounding rocks. The chilled margin would have more or less the same composition as the basaltic liquid.

But now we have a large column of hot basaltic magma, well insulated from the outside air. Crystallization of this large volume will continue over a very long period. The details of the crystallization process are complex, and we need not examine them in detail, but an important principle emerges from a brief examination of the process. Just as the first melt to be produced on partial melting does *not* have the same composition as the original rock, so the first crystals to form from a melt do *not* have the same composition as the melt itself. The process of crystallization therefore offers another way of making rocks with different compositions. Consider the data in Table 10.

The small amount of olivine that will crystallize from a basaltic magma is important to us in the present context because when the magma begins to crystallize, the *first* crystals to form are olivine, as you will appreciate from Figure 4 (page 10).

Table 10 shows that olivine contains silica, magnesium and iron. When crystals of olivine form, these components are removed from the melt.

☐ What, therefore, is the effect of crystallizing olivine alone on the composition of the melt?

■ Clearly, unless the mineral(s) removed from the melt by partial crystallization have exactly the same chemistry as the melt, the relative

TABLE 10 Chemical compositions of basalt and olivine.

	Basalt mass %	Olivine mass %
SiO ₂	49	40
TiO ₂	1.4	
Al ₂ O ₃	16	
Fe ₂ O ₃	2.2	
FeO	7.2	10
MgO	8.5	50
CaO	11	
Na ₂ O	2.7	
K ₂ O	0.3	
others	1.6	
	99.9	100

TABLE 11 Ingredients remaining in 90 tonnes of olivine-depleted liquid basalt: (a) in tonnes and (b) as mass per cent.

	(a)	(b)
SiO ₂	45	50
TiO ₂	1.4	1.6
Al ₂ O ₃	16	18
Fe ₂ O ₃	2.2	2.4
FeO	6.2	6.9
MgO	3.5	3.9
CaO	11	12
Na ₂ O	2.7	3.0
K ₂ O	0.3	0.3
others	1.7	1.9
	90	100.0

abundances of the different atoms will change. In the case of iron and magnesium oxides, which are more abundant in olivine than in the melt, their melt concentrations will *decrease* as olivine crystals are formed. Silica has a higher initial concentration in the melt than in the olivine being formed, so the effect of olivine crystallization is to *increase* the silica concentration in the residual melt. The same thing happens, of course, with elements that are not even present in olivine (see Table 10)—their concentrations in the melt increase during crystallization.

These important relationships can be best understood by working through a particular case. Consider a mass of 100 tonnes of magma, from which ten tonnes of olivine have crystallized (we'll see what happens to the crystals in a minute). The ten tonnes of olivine contain four tonnes of silica (SiO₂), one tonne of iron oxide (FeO), and the equivalent of five tonnes of magnesium oxide (MgO). These ingredients are subtracted from the magma, and the 90 tonnes remaining therefore contain the masses of other elements that are shown in Table 11.

If you compare analysis (b) of the olivine-depleted liquid basalt in Table 11 with that of the basalt in Table 10, you will see that there are some subtle differences, which are illustrated in the histogram of Figure 7. You might be thinking at this point that subtle differences in columns of figures are only likely to be interesting to an accountant. They are, however, crucially important to geologists, because they reveal that two quite different rocks can be made, simply by subtracting crystals of olivine from a basalt magma. This process is known as **fractional crystallization**. A similar technique, fractional distillation, is used in the oil industry to separate various petroleum products from crude oil.

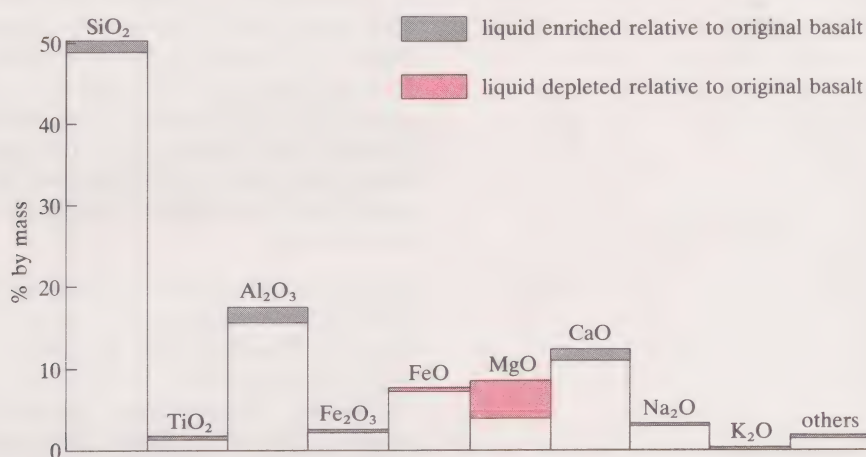


FIGURE 7 Histogram showing the differences in composition that are brought about by subtracting olivine from a basaltic magma.

In the case that we are considering, the olivine crystals that form in the magma may grow into substantial lumps. These are denser than the surrounding magma, and so sink to form layers, which accumulate towards the bottom of the pluton. These layers are effectively isolated chemically from the residual melt and form rocks known, not surprisingly, as **cumulates**. Since cumulates consist mainly of olivine crystals, they will have the composition mainly of olivine and will appear not unlike the peridotites from which we started. Cumulate rocks consisting mostly of olivine are common; more rarely, the cumulate minerals are oxides of iron, titanium and chromium, and these can be of major economic importance.

If the liquid left behind when olivine crystals have been removed could somehow be abruptly chilled, it would have the composition represented in column (b) of Table 11, which is effectively basalt minus olivine. But it is much more likely that crystallization will continue, in which case olivine will be joined by, and replaced by, other minerals, notably Mg-rich pyroxenes and Ca-rich feldspars (see Figure 4). These minerals will remove other elements from the liquid, so it will steadily depart from its original composition. When almost the whole of the pluton has crystallized, the last small

GABBRO

liquid fraction may have a composition that is vastly different from the original. In extreme circumstances, this tiny fraction may solidify to form a rock that is totally unlike basalt and approaching the composition of granite. (It is important to realize that *not* all granites are formed from basalt magma by fractional crystallization.)

ITQ 8 Look again at the analyses of basalt, the intermediate rock type andesite, and granite in Table 4. Can you see any trends in the proportions of the elements in each? Can you suggest what significance this might have in terms of fractionating crystals?

It is implicit from the answer to ITQ 8 that there is a continuous spectrum of rock compositions between basalt and granite, characterized in particular by increasing SiO₂ content. Fractional crystallization and partial melting are two ways in which rocks of different compositions can be made, but it is important to grasp that *both* these processes operate in the same direction, towards increasing silica content or, in other words, towards rocks of increasingly granitic nature. Thus, whereas minerals that melt at the lowest temperatures in a basalt or a gabbro could be partially melted to make a small amount of granite, a basaltic liquid can also be fractionally crystallized to make a little granite. In both cases the amount of granite that can be produced is very small. More usually, larger quantities of intermediate rocks are generated by both partial melting and fractionation of basaltic magma; the generation of larger quantities of granite requires several episodes of chemical distillation each of which progressively 'refines' the granitic product.

In both partial melting and fractional crystallization, there are some rocks that are left behind. In the case of partial melting of mantle peridotite, what remains is a large volume of depleted peridotite, from which the basalt has been squeezed out. This remains in the mantle, and is rarely seen at the surface. Similarly, fractional crystallization produces cumulates as well as silica-enriched liquids. The cumulates, necessarily, tend to be formed at deep levels, and are therefore not so frequently encountered as the rocks made from the remaining liquids. It is these that we deal with in the next two Sections.

One final point. We have been discussing here what happens to the composition of basaltic liquids if crystals are *removed* from the liquid by, for example, the settling out of crystals. If crystals cannot be removed—if, for example, the liquid is too stiff or viscous to allow them to settle—then, clearly, the composition of the rock that is ultimately formed will be the same as that of the basalt. Such rocks, of course, will be coarse-grained basalts or **gabbros**. These exist in huge quantities beneath the basalts of the oceanic crust.

3.5 DESTRUCTIVE PLATE MARGINS AND THE ORIGIN OF ANDESITES

You were introduced in Units 7–8 to the volcanic processes that accompany the descent of oceanic lithosphere into the mantle at a destructive plate margin. In this Section, we are going to look more closely at the processes that take place there. Consider first what happens in the simplest case—where one slab of oceanic lithosphere is forced beneath another.

- ☐ Can you recall what kind of surface features characterize such a subduction zone?
- ☒ Along the subduction zone at the point where the oceanic lithosphere descends is a marine trench behind which a line of volcanoes forms an *island arc*.

The significance of these island arcs is that they are located in the oceans, so there is no question of there being any continental crust beneath them.

Now, when a slab of oceanic lithosphere is forced down along a subduction zone, there is a great deal of friction between the descending and overriding plates. As you have already seen, this friction is manifested in the form of intense seismic activity and the liberation of heat.

As the descending slab of oceanic lithosphere is thrust deep down into the hot mantle, it is also heated by conduction. The heat produced by friction and conduction ultimately leads to partial melting and the generation of magmas, which are erupted by the volcanoes along island arcs.

So sufficient heat is generated to cause partial melting. But what is it that melts?

It is possible that what melts is some of the rocks of the oceanic plate and/or those of the overriding mantle against which the plate is grinding.

- ☐ Whichever rocks melt, would you expect the magma to have the *same* composition as the melted rocks?
- ☒ No, you should recall from Section 3.3 that when rocks are melted, the first melts that are produced do *not* have the same composition as the melting rock.

Within the oceanic lithosphere it is the uppermost rocks—the basalts and gabbros—that have the lowest melting temperature and which are, therefore, likely to melt first. But partial melting of basaltic rocks of the oceanic plate would *not* produce a new generation of basaltic magma.

Unfortunately, life is not quite that simple. Beneath *constructive* margins, new mantle comes towards the surface and basaltic magma is extracted from it. The depleted mantle then moves on and diverges outwards forming the lower parts of the oceanic lithospheric plate. The next bit of mantle in the conveyor system comes up and goes through the same process. On a gross scale, one bit of mantle has the same composition as the next, so the new oceanic crust produced from it will have the same overall composition.

At *destructive* margins, by contrast, although the down-going plate moves continually, the overlying plate stays in the same position with respect to the subduction zone beneath it (Figure 8). This being so, magma is continuously produced at the same site, relative to the overriding plate. As time

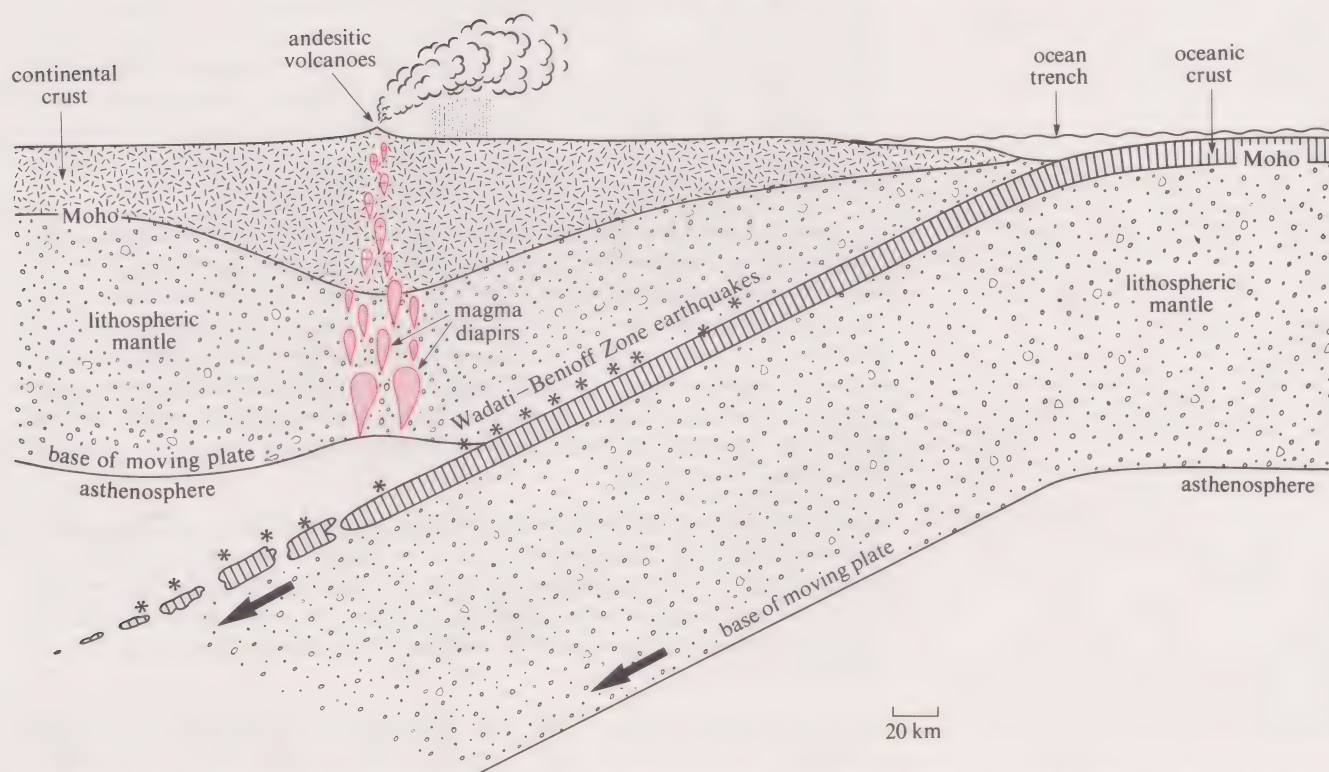


FIGURE 8 Subduction zone processes. New parts on the descending oceanic plate are constantly undergoing partial melting, but the magmas produced continue to rise through the same parts of the continental crust. In this diagram, the components are drawn to scale, as far as possible.

ANDESITES

passes, more and more melts will pass upwards through the *same* zones in the overlying plate, which will be progressively heated up and will change in composition as magma is either added or subtracted. In these circumstances, it follows that the first magma will differ from those developed later on. In areas where we know that subduction is new and volcanic activity commenced recently, the first lavas are basaltic and similar in composition to those of the oceanic crust. The later products are quite different from those primitive basalts, and are much more siliceous. Thus it seems that partial melting of mantle peridotite may be involved at an early stage but that more complex processes—perhaps involving the remelting of basaltic ocean crust as the island arcs become more mature and more siliceous magmas are produced—are also involved.

The resulting siliceous rocks, which are characteristic of destructive plate margin volcanoes, are known as **andesites** (Units 7–8). The name, of course, is derived from the Andes, where such rocks are very abundant. They are generally lighter in colour than basalts and slightly lower in density.

- ☐ What do the lighter colour and lower density suggest to you about the minerals present in andesites?
- These facts indicate that andesites contain *fewer* ferromagnesian minerals than basalts.

Andesites have chemical compositions between those of basaltic and granitic rocks (Table 4), and many of their physical properties, such as colour and density, are also intermediate.

There is one crucial difference between andesites and basalts. Whereas most basalts are created from mantle rocks at constructive plate margins, and return to the mantle at destructive plate margins, the manufacture of andesites is a one-way process; andesites do not return to the mantle. Thus, when a volcano at a subduction zone erupts, it not only provides a firework display of lavas and ashes but also yields a virtually *irreversible addition to the continental crust*. The volcanoes of the Andes, therefore, are characterized by steep-sided cones formed of layers of pyroclastic ashes interspersed with pumice and rock debris. The magmas are very viscous and solidify more readily in the vent of the volcano. With the main exit blocked, pressure builds up within the volcano. Often this pressure is released by a massive eruption that blows part of the top or side of the volcano away. The eruption of Vesuvius in AD 79 and the 1980 eruption of Mount St Helens (see cover illustration) were of this type. Satellite photographs show that much of the Andes is composed of explosive cones superimposed on each other. One of the most dangerous aspects of these volcanoes is their unpredictability compared with those created from basaltic magmas.

In Units 5–6 you saw that, broadly speaking, the continental crust is composed of a low-density igneous rock, between granite and basalt in composition, and a veneer of sediments. The rock that is intermediate between granite and basalt is, of course, andesite (and its plutonic equivalent diorite), and it will be argued in the next Section that the continents themselves originated largely through andesitic volcanism at destructive plate margins. The terminology is perhaps a little confusing here: *oceanic* crust is created at *constructive* plate margins, but *continental* crust is created at *destructive* plate margins!

Consider for a moment a hypothetical Earth in which there are no continents, but there is a hot, mobile mantle in which convection currents are active. Plate tectonic processes of a sort will operate (they do not require the presence of continents) and, in particular, there will be a number of destructive plate margins at which andesites are being erupted at island arc volcanoes.

With the passage of time, some arcs will be rapidly eroded, and may even disappear below sea-level. But, in general, the material tends not to disappear into the mantle—for one thing, its density is too low. Along the site of

the inactive island arc, then, there will be a long narrow belt of rocks of andesitic composition, mostly sediments, but with some of the original lavas intact. This inactive belt will drift around passively on the surface of the Earth for tens or hundreds of millions of years. Eventually, however, plate tectonic processes are bound to lead to its colliding with *another* arc, possibly an active one (Figure 9).

Because of its low density, the ancient island arc cannot be forced down into the mantle beneath the new arc—it would be a bit like trying to push a beachball under water. The net result is that the old island arc collides with and accretes to the new one, subduction ceases, and the two become welded together to form a mini-continent. Moreover, with the evolution of the island arc, this mini-continent will have become increasingly andesitic, and will, therefore, be approaching continental crust in composition. With the continuation of plate tectonic processes, it is easy to see that as more and more island arcs are accreted together, the size of the accreted mass will increase until it reaches true continental dimensions. Once the destructive margin has evolved to the extent that subduction is occurring beneath the continental crust, it is no longer an island arc but has become an Andean destructive margin.

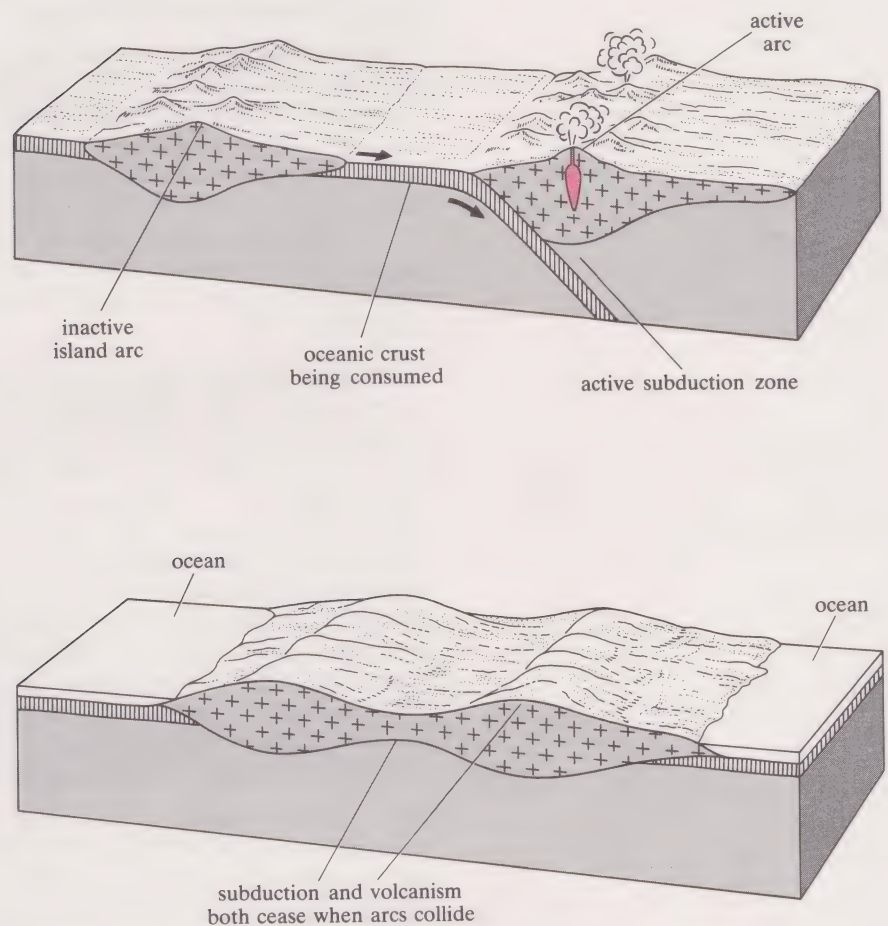


FIGURE 9 The accretion of island arcs to form a mini-continent.

GRANITE

DIAPIRS

3.6 GRANITES AND THE ORIGIN OF CONTINENTS

You will by now be familiar with the **granite** sample S1 in your Kit, and you will also have determined its density. Your sample is typical of plutonic igneous rocks: it is coarsely crystalline, the crystals are tightly interlocked with one another, and the minerals present are easily recognizable.

Whereas basalts are erupted at temperatures of over 1100°C and flow freely like ordinary liquids, the silicate minerals of which granite is made melt at *much* lower temperatures (Figure 4), and the viscosity of granite melts is much higher. It is thought, in fact, that a granitic magma is more like a pasty mush than a free-flowing liquid; hence, when crystals form in it, they do not settle out but grow larger where they are, eventually interlocking with other crystals.

When you determined the density of your granite specimen, you should have found that it was lower than that of any of the other igneous rocks. Consider a granitic magma of density $2.4 \times 10^3 \text{ kg m}^{-3}$, which is formed at the base of continental crust with density $2.9 \times 10^3 \text{ kg m}^{-3}$. Since it is less dense than the surrounding rocks, it will tend to 'float' up to the surface. Granitic magmas, therefore, have a powerful upward urge and this enables them to rise great distances through the crust.

If such a rising granitic body enters a region of rocks of lower density, it will cease to rise; similarly, if it cools and solidifies, it will grind to a halt. Until either of these two things happens, granites will rise through the crust, pushing aside the surrounding rocks. In some cases they come extremely close to the surface, to within a matter of one or two kilometres. In other cases, they come so close to the surface that a conduit is blasted through to the open air, and magma makes its way to the surface forming a volcano. In this case a fine-grained *volcanic* rock is formed, with the same chemical and mineral composition as granite; this is called rhyolite (and is shown in Plate 19 of Units 5–6).

Granites are found in large volumes at destructive plate margins, especially in areas such as the Andes where the continental crust is very thick. But granites form only a few per cent of the dominantly andesitic continental crust. How then do these granites form? To answer this, we need to look more closely at the igneous process taking place at destructive plate margins. We have said that, in the simplest case of an island arc, basaltic and andesitic magmas are produced by partial melting of mantle material and subducted ocean crust. As you know, small amounts of granite can be formed by distilling a basaltic magma through partial melting and fractional crystallization. But it is believed that most granites are not formed directly from basalts; in a sense, they are refined andesites, resulting either from fractional crystallization of andesitic magma or from partial melting of andesitic crust. Of course, a larger volume of granitic melt can be produced from this intermediate-composition parent than from basalt.

What seems to happen is that in its early stages, the overriding plate at a subduction zone is relatively cold and brittle. Andesitic magmas that form early aggregate into blobs or **diapirs** (shown in Figure 8), which then rise into the upper crust. Initially, the melts will not rise very far before they solidify. However, the process is continuous, and as more and more diapirs rise, so the overriding plate becomes hotter and hotter. Accompanying the eruption of unchanged andesitic magmas, partial melting of some of the early-formed andesitic rocks takes place, and this gives rise to the more siliceous magmas. These then rise to higher levels in the crust before solidifying to form coarse-grained granitic rocks.

In this way, at an island arc where initially nothing but oceanic crustal rocks are present, increasingly siliceous crust is generated, which becomes progressively more continental in character. This process seems to take place relatively quickly because, in areas where island arcs are no more

than 20 million years old, continental crust up to 30 kilometres thick has been identified by seismic studies. It is also possible that most of the oldest continental rocks, older than 2 600 million years, were formed in this way.

SUMMARY OF SECTION 3

1 80% of the Earth's volume and about 70% of its mass consists of mantle rocks with the composition of peridotite. Mantle peridotite does not, however, contain quite enough radioactive isotopes to be the sole source of the Earth's internal energy. There is a significant contribution from crustal rocks and it is also possible that some primordial heat derived from the formation of the Earth may still make a small contribution to surface heat flow.

2 Volcanic rocks crystallize at the Earth's surface and so are usually fine grained; they rest on top of older rocks.

3 Basalts are characteristic of constructive plate margins, where they are almost certainly formed by partial melting of mantle peridotite. Partial melting occurs when the minerals with the lowest melting temperatures (pyroxene and plagioclase feldspar in peridotite) start to melt and form a magma that, after crystallization, produces a different rock type, basalt. High-temperature, low-viscosity basalts spread out into broad, gently sloping shield volcanoes.

4 Plutonic rocks are formed underground. They are coarse grained and cut across older rocks.

5 Fractional crystallization is the process by which the minerals crystallizing out from a melt leave a residual liquid of a different composition. In basalts, the first crystallizing mineral is often olivine, which may form cumulates within a magma chamber. The residual liquids produced by fractionation of basalt are usually of intermediate composition but they can, eventually, reach the composition of granite.

6 Partial melting of the basalt layer in the downgoing oceanic lithosphere beneath island arcs and active continental margins produces andesitic magma. This magma rises and is intruded as diorite or is erupted, forming steep-sided andesitic volcanoes. The morphology of these volcanoes is influenced by the higher viscosity of andesite compared with basalt.

7 The net result of all the processes taking place at subduction zones is that throughout geological time there has been a slow and steady increase in the volume of andesitic continental crust at the expense of mantle peridotite. This is an extremely important result, so it is worth summarizing the main steps:

(a) Partial melting of mantle peridotite and basaltic ocean crust beneath destructive plate margins produces basaltic and andesitic magmas.

(b) Andesites accumulate almost irreversibly to form island arcs.

(c) Prolonged magmatism beneath island arcs leads to re-melting and to progressively more siliceous magmas, which may eventually reach granitic composition.

(d) Island arcs accrete to form mini-continents.

(e) Continuing destructive plate margin activity adds siliceous magmas, thus increasing the thickness of the continental crust and enlarging the mini-continent.

8 Magmas of granitic composition are formed mainly by partial melting of andesitic material at destructive plate margins. Being less dense they rise through the overlying crustal rocks before solidifying into coarse-grained plutonic rocks. Rarely, magmas of granitic composition reach the Earth's surface, where they may form lavas called rhyolite.

WEATHERING

PHYSICAL WEATHERING

CHEMICAL WEATHERING

SAQ 3 Which of the statements below correctly describe the source of the Earth's internal energy? Give a brief statement of your reasons.

- A Most of the Earth's internal energy comes from radioactive decay of elements in the Earth's crust.
- B The rate of radioactive heat production within the Earth must always have been the same as it is today.
- C The Earth may retain within it some of the heat generated during its formation 4.6 billion years ago.

SAQ 4 Which elements would be relatively depleted in the residual liquid that is left when plagioclase crystals consisting of 100% (i.e. pure) $\text{CaAl}_2\text{Si}_2\text{O}_8$ are removed from an andesitic magma? Which elements would show the greatest relative depletion? Explain your answer. (You may assume that the composition of the plagioclase is (by mass): 57% SiO_2 ; 30% Al_2O_3 ; 13% CaO .)

SAQ 5 Which of the following characteristics apply *only* to volcanic rocks? (Some items may be irrelevant.)

- A formed by fractional crystallization
- B formed underground
- C older than surrounding rocks
- D coarse-grained margins
- E fine-grained margins
- F coarse-grained all through
- G fine-grained all through
- H rest on top of younger rocks
- I rest on top of older rocks
- J formed by partial melting

SAQ 6 There are 50 000 kilometres of active destructive plate margins in the world. Assuming that each kilometre contributes 25 km^3 of magma to the surface per million years, and that 80% of this is derived from the mantle, how much *new* continental crust is generated each year?

SAQ 7 Where do you suppose the 20% of the magma not accounted for in SAQ 6, i.e. not derived from the mantle, comes from?

SAQ 8 The vast majority of granites in the world are formed at destructive plate margins by partial melting of older continental rocks. By what other mechanism could small volumes of granitic magma be made?

SAQ 9 Fill in the missing words in the following paragraph from those in the key.

Granites may be formed in small volumes at plate margins by extreme crystallization of magmas, but they are formed in much larger volumes at plate margins, where progressive episodes of magmatism lead to the evolution of progressively more liquids. Heat carried up into the crust may also cause partial in the upper crust, thus contributing further to the volume of granitic magma.

KEY

basaltic	constructive	destructive
melting	siliceous	fractional

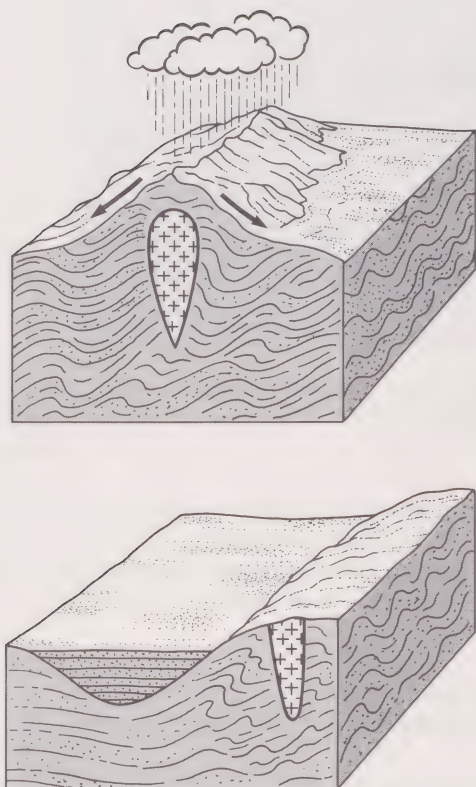
4 SEDIMENTS AND SEDIMENTARY ROCKS

4.1 INTRODUCTION

In going on to look at the ways in which sediments and sedimentary rocks are formed, we are moving on to the next stage in the rock cycle, shown in the second part of Figure 2. This is the stage that takes over once the processes of igneous rock formation are complete and the igneous rocks are exposed at the Earth's surface, either because the magmas were erupted initially as lavas or because the rocks into which they were intruded have been worn away. Such igneous rocks are exposed to a totally different environment at the surface of the Earth from the one in which their constituent minerals crystallized. They are in disequilibrium with these conditions: temperature and pressure are both much lower, and they now have an abundant supply of water, which is oxygen rich.

In response to these environmental differences, many of the silicate minerals of igneous rocks undergo changes that may lead to their *total or partial breakdown* and, consequently, to the destruction of the rocks themselves. These breakdown processes are known collectively as **weathering**, and it is the products of weathering that eventually accumulate to form sedimentary rocks. This is one part of the rock cycle that you can see for yourself in all parts of the country. However, you should not get the idea that it is only igneous rocks that are subject to weathering. As you will see in Section 5.4, metamorphic processes lead to the formation of rocks at a wide range of temperatures and pressures (encompassing, at the upper limit, those of igneous rock formation), so they, too, are vulnerable to attack by weathering when subjected to the normal temperatures and pressures at the Earth's surface. Even sedimentary rocks themselves may be subjected to fresh cycles of weathering.

The TV programme 'From Snowdon to the sea' is particularly relevant to this Section. The notes that accompany it are in Section 4.8.



4.2 PHYSICAL AND CHEMICAL WEATHERING OF ROCKS

Two distinct types of weathering can readily be distinguished: **physical weathering** (or mechanical disaggregation) and **chemical weathering** (or chemical decomposition). Both processes, either separately or together, lead to the formation of a wide range of sedimentary rocks.

One of the most important forms of physical weathering, which causes the break up of large masses of rock into fragments, results from the repeated alternation of freezing and thawing of rainwater in cracks and fissures. Unlike most other substances, water *expands* on freezing, with an increase in volume of about 9%, and the resulting pressure increase on the sides of the fissures causes them to split further apart. Large blocks of rocks fall under gravity from rock faces and are shattered into smaller angular fragments, which accumulate as scree slopes. This process is called *frost shattering* and is responsible for the splitting of rocks into the broken slabby boulders shown in Figure 10. Shattering of rocks into fragments like this exposes a much larger surface area to rainwater run-off, and so the fragments may soon undergo chemical weathering. This attack is enhanced by the presence of dissolved atmospheric gases such as carbon dioxide (and also sulphur dioxide) in the rainwater, which make it slightly acidic. Chemical reactions may take place between the rock minerals and water, sometimes dissolving them completely and sometimes producing new mineral types.

RESIDUAL MINERALS

CLAY MINERALS



FIGURE 10 Frost shattered boulders. The cracks and crevices in which water collects can be seen clearly in the boulders, and the various stages in the break-up of the boulders are visible.

You will recall from Section 2.3 that the silicate minerals of igneous rocks crystallize in a sequence that is controlled by their temperature of formation. Each of these minerals varies in its susceptibility to chemical attack, a property that is described as the *relative resistance of the mineral to chemical weathering*.

- Look at Table 12, in which common minerals are listed in order of their relative resistance. How does this compare with the order of crystallization shown in Figure 4?
- The first-formed, high-temperature silicate minerals are the first ones to break down, whereas the last-formed, low-temperature minerals are the most stable.

TABLE 12 The relative resistance of silicate minerals to weathering

	Ferromagnesian minerals	Iron- and magnesium-poor minerals
Increasing resistance to chemical weathering ↓	olivines pyroxenes amphiboles biotite mica	calcium- and sodium-plagioclase feldspars potassium feldspar muscovite mica quartz

Note: you have seen all these minerals in the Earth sciences laboratory at Summer School.

In fact, rainwater has virtually no chemical effect on quartz at all. This sequence of silicate mineral stability is not coincidental: it is directly related to the chemical compositions and structures of the minerals themselves. In simple terms, silicon ions bond strongly to oxygen, to form silicate anion complexes, mainly because they are small and highly charged. However, the other cations in silicate minerals are both larger and more weakly charged, so bonds between silicate anion complexes and the other cations present are relatively weak. So, minerals with weak cation–silicate bonds, such as olivine, pyroxenes and Ca–Na feldspar, are more easily broken down than those with mainly silicon–oxygen bonds, such as quartz.

4.3 THE PRODUCTS OF CHEMICAL WEATHERING

When a rock is exposed to chemical weathering, the following things happen.

- (a) Substances that are soluble in water are carried away in solution by rainwater and rivers, and eventually enter the sea.
- (b) New minerals are formed when soluble substances are leached out of existing minerals and their atomic structures partially collapse. An example of this process is the formation of clay minerals through the weathering of feldspars and micas in granite.
- (c) Resistant minerals, such as quartz, are not chemically attacked at all.

The weathering products of average igneous rocks are shown in Figure 11.

The new minerals and the resistant minerals are known as **residual minerals**. The most common of the resistant minerals is quartz but, if chemical weathering is not too prolonged, feldspar and micas will also resist chemical attack. Quartz is found in huge volumes as a residual mineral in the form of sand grains on beaches all over the world (although not all beaches are made up of quartz sand grains).

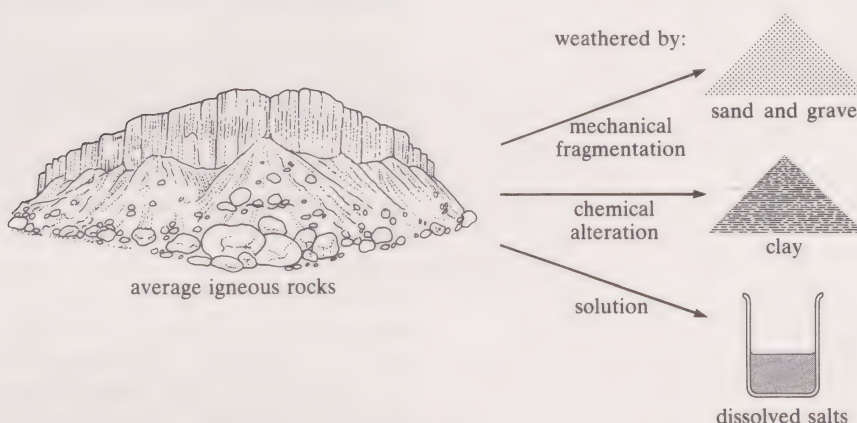


FIGURE 11 The weathering products of average igneous rocks.

The breakdown of ferromagnesian minerals, such as olivines, allows magnesium to be carried away in solution, but this usually leaves a residue of insoluble iron oxides and hydroxides, which gives some weathered rock surfaces their rusty brown colour. Silica, in the form of $(\text{SiO}_4)^{4-}$, from the silicate minerals, is also very slightly soluble. When the soluble cations are leached out of the more complex silicates (such as feldspar where the silicate framework forms large molecules), the collapsed atomic structure usually incorporates water to form a flaky, hydrated **clay mineral**.

There is more than just academic interest in the breakdown of feldspar to clay minerals. In Cornwall, for example, the feldspars present in vast quantities of granite have been broken down by complex chemical and physical processes to form valuable deposits of *china clay*, which have been worked commercially for centuries (Figure 12). China clay has many uses; besides being used in the pottery industry it is found in pharmaceuticals as a 'carrier' for drugs in pill form, and also as a 'filler' in paper to make the surface smooth and glossy. There are several grams of china clay in the very Unit you are reading!

ITQ 9 What would you predict are the main products of intense chemical weathering of the following types of rock?

- (a) granite (dominantly feldspar with smaller quantities of mica and quartz);
- (b) gabbro (dominantly feldspar and pyroxene with, sometimes, a small amount of olivine);
- (c) quartz sandstone with the grains cemented together by calcium carbonate.

 EROSION

 DENUDATION

 BEDLOAD

 SUSPENDED LOAD

 ENERGY OF THE
ENVIRONMENT



FIGURE 12 An aerial photograph of the spoil heaps associated with the china clay industry in Cornwall. These heaps consist mainly of weathered rock fragments, quartz and some residual clay and mica.

As you might expect from this discussion, climate plays a very important part in weathering. Very little weathering of any sort can take place unless there is at least a little rainfall each year. For chemical weathering to be most effective, annual rainfall should exceed about 1000 mm and temperatures should be high, to speed the rate of the chemical reaction. Frost shattering can still be very effective at quite low annual rainfalls, but obviously it requires the temperature to fluctuate above and below freezing. For these reasons chemical weathering predominates in the hot, humid low latitudes and physical weathering predominates in the colder high latitudes. Here in the UK our climate lies somewhere between these extremes.

- ☐ Try to think of two sets of climatic conditions in which chemical weathering would be insignificant, or of reduced significance, at low latitudes
- In hot desert regions, such as the Sahara Desert, rainfall is extremely rare and virtually no chemical weathering occurs between rainstorms. In mountainous regions, temperatures decrease with altitude and so mountainous areas in low-latitude regions often experience much more physical weathering than the surrounding low-altitude countryside.

We must stress that not all chemical weathering takes place at the site of a rock outcrop being weathered. As soon as rocks have become fragmented the fragments are moved, under gravity, by water, wind or glaciers, if climatic conditions are right. So both physical and chemical breakdown continue during transportation of the fragments.

4.4 THE TRANSPORTATION OF WEATHERED MATERIAL

It is fairly obvious that the soluble products of chemical weathering are removed by rainwater and rivers. However, the residual products of chemical weathering, and the larger fragments produced by physical weathering, are also removed by various transporting media. The removal of weathered material is usually referred to as **erosion**, and the combined processes of weathering and erosion that gradually wear down landscapes are termed **denudation**.

Wind and ice, as well as water, can be important transporting media, although fortunately they are not particularly significant in Britain at the present day. However, as you saw in TV 7, there is evidence in our geological record to show that there have been periods of time when Britain was subjected to wind erosion in desert conditions, and scoured by glaciers and ice sheets. As it is principally water movement, either by river or the sea, that concerns us in Britain today, we shall concentrate on this.

The transport of rock fragments and mineral grains by water involves either the bouncing or rolling of the material along the bed of a river or the sea as **bedload**, or the transport of the material within the water itself as **suspended load**. If you took a spadeful of sediment and dropped it into a moving stream, whether it was transported as bedload or suspended load would depend on two factors: the size of the particles and how fast the stream was flowing. The coarser material would tend to move as bedload, and only the finer sediment would be carried along within the water itself as suspended load. The largest fragments of all might well be deposited and remain immobile on the stream bed.

- ☐ If you repeated this experiment the following day when the stream was flowing much faster because of heavy rainfall overnight, would you expect the same results?
- ☒ In the broadest terms, yes. But now you would find that some of the coarser material previously carried as bedload had been lifted into suspension by the faster moving, higher energy water. This is important because it means that it is the **energy of the environment** that determines how a sedimentary particle is transported and, indeed, whether it is transported at all.

Water flowing at a given, constant speed will transport all particles up to a certain maximum grain size (expressed in terms of the diameter of the grain), either as bedload or as suspended load. If the speed decreases then some of the suspended load will fall to the bottom and move as bedload and some of the bedload will stop moving altogether; in other words, it will be *deposited*. The grain sizes that will be deposited at various current speeds are determined by the lower line in Figure 13. Although coarser sand and gravel carried in suspension by fast-moving water have to pass through a

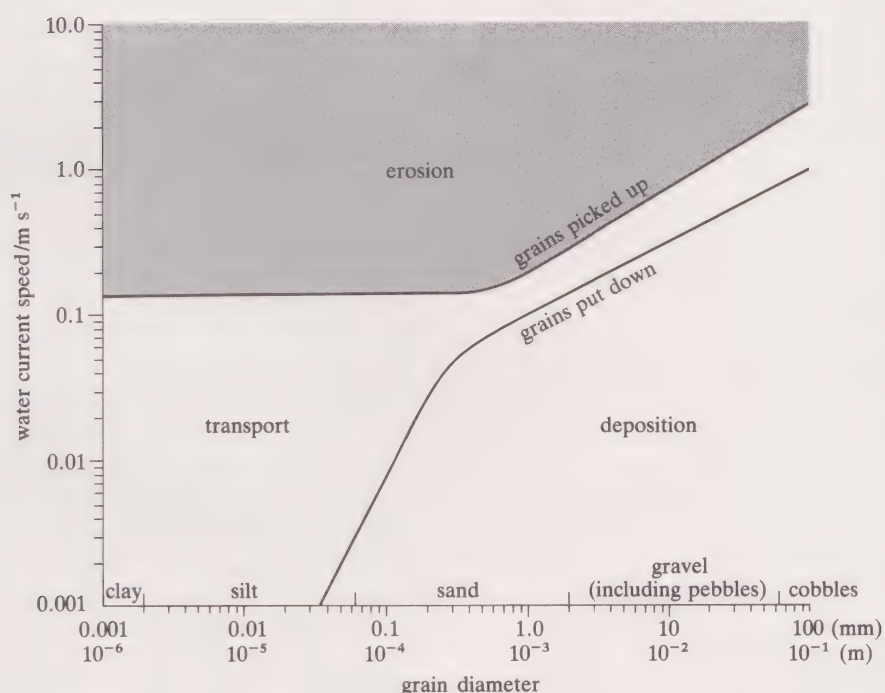


FIGURE 13 Graph to illustrate the relationship between the size of sediment grains and the speed of a water current necessary to pick them up (upper curve) and deposit them (lower curve). The zone between the two lines represents the field in which sediment grains can be carried in suspension by the current, once they have been picked up.

SORTING

phase of bedload transport before they are deposited, finer sand and silt are deposited as soon as they fall out of suspension. This has important consequences for the nature of sedimentary deposits as you will see later (Section 4.5).

Water current speeds and changes in speed are largely responsible for the selective uptake and deposition of sediment grains, or the process of sorting a sediment, as well as whether, overall, it is coarse or fine grained. The **sorting** is simply a measure of the range of grain sizes about the mean grain size value for that deposit. A well sorted sediment has grains that are very similar in size and that have a narrow range about the mean, whereas a poorly sorted sediment has grains of various sizes with a wide range about the mean.

ITQ 10 Using Figure 13, answer the following questions.

- What is the maximum grain size that can be moved by a vigorously flowing stream with a water current flowing at 0.3 m s^{-1} ? How will such grains be moving?
- Will the sediment resting (i.e. deposited) on the bed beneath the water be coarse or fine?
- What is the water current speed required to deposit a very fine sand with grains of 0.1 mm diameter?

It is quite clear from the answer to ITQ 10 that high-energy water environments result in the deposition of coarse-grained sediments, and low-energy environments lead to the deposition of fine-grained sediments. There can be a good deal of variation in the type of sedimentary rock within a small area. For example, around headlands where wave action is concentrated, large boulders and pebbles are strewn on the beach, whereas in the sheltered bays, sands are deposited (Figure 14). Similarly, on the outside of a river bend (or meander), where the flow is fastest, coarse sands and gravel may be found on the river bed whereas on the inside of the bend, where the flow is slowest, finer sand may be deposited.

- ☐ Grains of silt-size, or fines, are only deposited when there is almost no water movement at all. From general knowledge, can you suggest any environment where this condition might be fulfilled?
- ☒ Suitable environments include lakes; river flood plains after flooding has occurred and the river has returned to its normal channel, leaving vast pools of water on the flood plain; the parts of a river estuary or coastal mud flat that are covered by water only at the slack waters of high tide; and the sea-bed below the influence of normal wave and tidal-current action.

The sorting of sedimentary material is a progressive process. The further a mass of sedimentary material is transported, the greater is the chance that it will become separated out into fractions of different sizes. Sometimes this separation is aided actively by the way in which water moves. For example, the constant action of waves on a beach winnows out the finer sands and silts and keeps them in suspension to be deposited later where low-energy conditions prevail, such as in a sheltered bay or the mouth of an estuary.

Much of our discussion so far has assumed that the particles being transported are all of the same mineral type. But the mineral type is important because it determines the shape of the sediment grains and their density. Flaky minerals, such as micas, are not deposited according to the conditions shown in Figure 13 because their shape offers considerable resistance to the water as they settled. This is why mica flakes tend to occur in sandstones and siltstones that have been deposited in water under low-energy conditions. In fact, Figure 13 is based on the behaviour of spheres of quartz, which are the most common mineral grains transported. As you know, density varies from one mineral to another. The density of quartz ($2.65 \times 10^3 \text{ kg m}^{-3}$) is close to that of feldspar ($2.6\text{--}2.7 \times 10^3 \text{ kg m}^{-3}$), and



(a)



(b)

FIGURE 14 (a) Large boulders at the foot of headland cliffs, where the energy and wave action are high. (b) Small particles of sand in a bay where the energy and wave action are low.

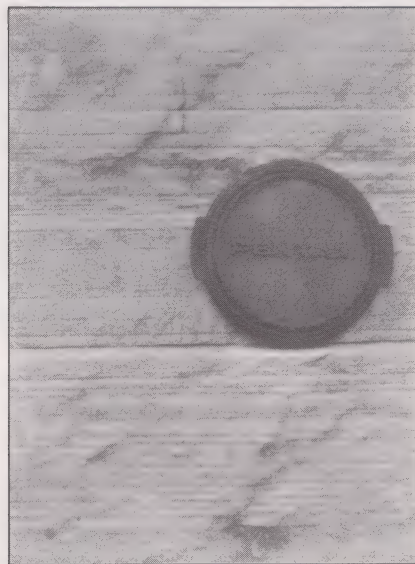
since these are two of the most important rock forming minerals, it follows that the densities of most common silicate rocks fall within a fairly narrow range (e.g. gabbro, $3.0 \times 10^3 \text{ kg m}^{-3}$ and granite, $2.4\text{--}2.6 \times 10^3 \text{ kg m}^{-3}$). However, there are some minerals that are *much* denser than quartz. Often these are minerals rich in metals, such as chromite (chromium rich, density $4.5\text{--}4.8 \times 10^3 \text{ kg m}^{-3}$), pyrite (iron rich, density $5.0 \times 10^3 \text{ kg m}^{-3}$) or cassiterite (tin rich, density $7.0 \times 10^3 \text{ kg m}^{-3}$).

Particles that have been picked up and transported are subjected to mechanical abrasion, and undergo changes in shape. It is fairly obvious that the longer particles are in motion, the more eroded they will become, as edges and corners are worn away. It is perhaps less obvious that *large* particles are much more rapidly rounded than small ones; this is mainly because large particles have more momentum than smaller ones and, hence, can do more damage to each other when they collide. The upshot of this is that rounded particles are much more common in coarse-grained sediments deposited in a high-energy environment than in fine-grained sediments deposited in a low-energy environment. This is something you can easily confirm for yourself on a beach that has both pebbly and sandy parts. The pebbles are almost always smoothly rounded, but the sand grains are much more angular.

4.5 BED FORMS AND SEDIMENTARY STRUCTURES

In Section 4.4 we said that whether sediment was deposited directly from suspension (as is the case with very fine grains) or went through a phase of bedload transport first (as with coarser grains) had important consequences for the nature of the final sedimentary deposit. Fine grains, falling out of suspension like confetti, come to rest on the bed beneath the water as very fine flat and even layers, or laminae, usually only a millimetre or so in thickness. So, fine-grained, *laminated sediments* are characteristic of suspension deposits (Figure 15a).

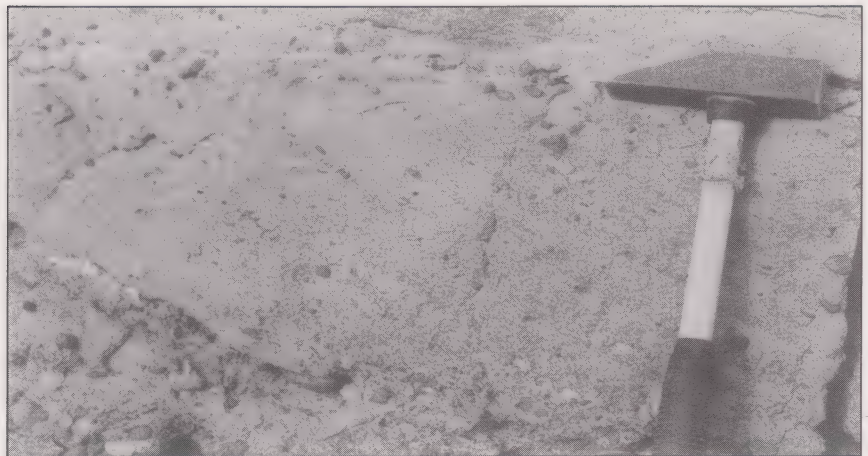
Coarser grained sediments, which move in the bedload prior to final deposition, do not form such neat layers. Often they are built up by the motion of the water near the bed into regularly spaced mounds and hollows known as bed forms. You are probably most familiar with these in the shape of the ripple marks that can be seen on many beaches and estuaries at low tide (Figure 15b). Once loose sediments have become compacted and cemented to form sedimentary rocks, rippled bed forms are preserved, characteristically as the series of inclined bedding planes known as *cross-bedding* that you first met in TV 7 (Figure 15c and Plate 12). The way these cross-beds develop is shown in Figure 16, where the current is flowing from left to right, sweeping particles over the ridge of the ripple to deposit them in the trough.



(a)



(b)



(c)

FIGURE 15 (a) Fine laminations in a sediment, characteristic of a suspension deposit. (b) Ripple marks, which can be found on many beaches and mudflats at low tide. (c) Small-scale cross-bedding in an ancient sediment, indicating a water-borne sediment.

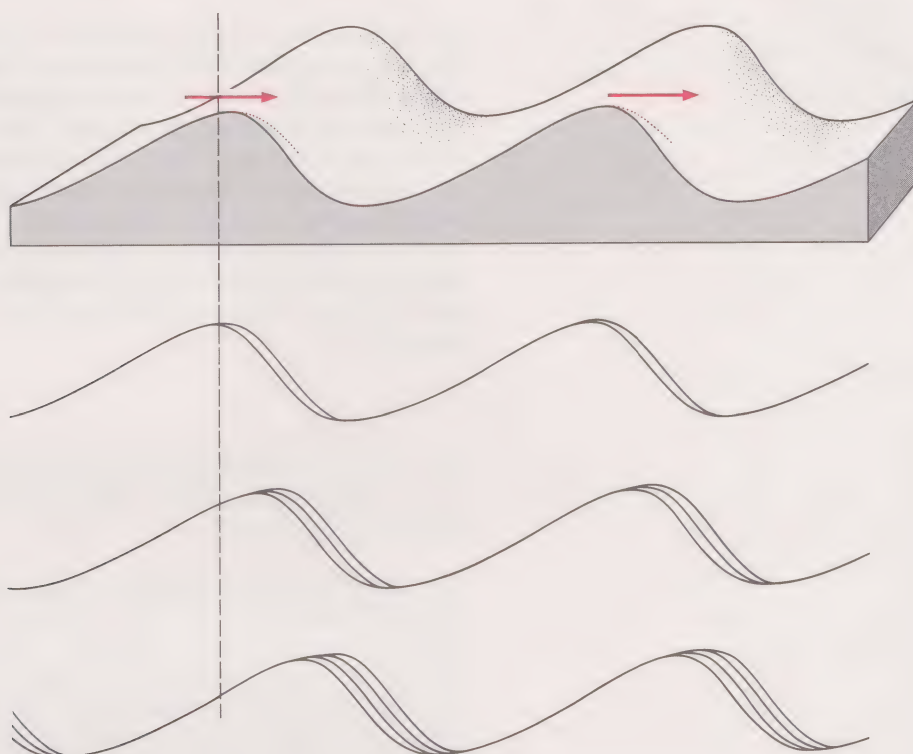


FIGURE 16 Stages in the development of cross-bedding. The current is moving from left to right, moving material over the crest of the ripple and depositing it on the leeward side. Eventually the material is buried by further advances of the ripple front.

4.6 THE TRANSPORT AND EROSION OF SEDIMENT BY WIND AND ICE

Although water is the most important medium for eroding and transporting sediment in Britain today, in other parts of the world and in previous periods of Britain's geological history, wind or ice have been more significant.

You may find it strange to consider that the air is a fluid, and that wind behaves in exactly the same way as water when transporting sediment. The only difference is that the density of air and, consequently, its viscosity (resistance to flow), is very much lower than that of water. Therefore, moving air can move a much more restricted range of grain sizes than water. To give you an example, whereas it takes a water current speed of only 0.2 m s^{-1} to start a quartz grain of 1 mm diameter rolling, it would take a wind speed of nearer 10 m s^{-1} (36 km h^{-1} or force 6 on the Beaufort wind scale) to do the same thing. Most transport by air is as a kind of bedload sweeping across the ground, as you may have seen in modern deserts, and only the very finest dusts are lifted up into suspension (Plate 13a). This means that, after re-deposition of the particles as wind speeds gradually fall, the resulting wind-blown sediments are usually very well sorted (Plate 13b). The individual quartz grains are usually more rounded than those of water transported sediments because the wind-blown grains collide without the cushioning effect of water, and so have their corners and edges worn down much more rapidly.

By contrast, moving ice, in the form of glaciers or ice sheets, is totally indiscriminate about what it transports. Anything from fine clay-sized particles to vast boulders several metres in diameter can be embedded in the ice. The mechanism of erosion is also different. Whereas some particles and even sizeable fragments may be picked up by the moving ice, many of the larger boulders fall from the sides of a valley onto the surface of a glacier. Subsequently, such boulders work down through the ice because of the greater density of the rock material. Once embedded in the ice, the fragments are not free to collide with each other, and most will not be deposited until the

LIMESTONE

OOZE

EVAPORITE

ice melts (though large boulders may be deposited at the base of the glacier), often many miles away from where the fragments were picked up. When the ice melts, all the remaining fragments, from clay-sized particles to boulders, are deposited at the same time to form a long ridge of sediment, known as a moraine (Plate 14). Consequently, true glacial sediments are very poorly sorted and the fragments are characteristically angular. The larger fragments are also often faceted, owing to scraping of the entrained boulders across the floor of a glaciated valley. This means that glacial sediments are not characterized by diagnostic grain shapes or degrees of sorting that will give you any clues about how far the material has been transported.

4.7 THE TRANSPORTATION AND DEPOSITION OF THE SOLUBLE PRODUCTS OF CHEMICAL WEATHERING

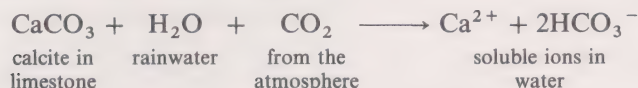
The most abundant soluble cations released from silicate minerals by chemical weathering are Ca^{2+} , Na^+ , K^+ and Mg^{2+} . These, along with some silica, in the form of $(\text{SiO}_4)^{4-}$, are carried away in rivers, eventually reaching the sea. Therefore, you might imagine that as weathered material in solution has been added to the oceans over a period spanning many millions of years, the sea should be becoming progressively more saline. This is not the case because the dissolved salts are removed at the same rate as they are added, which means that the composition of seawater has remained reasonably constant over at least the past 1 000 million years. Two major ways in which dissolved salts are removed are (i) by the action of marine organisms and (ii) by direct chemical precipitation.

(i) *The action of marine organisms*

Many marine animals build shells composed of calcium carbonate by extracting both Ca^{2+} and HCO_3^- from seawater to form the mineral calcite (CaCO_3 ; mineral sample IV in the Kit). After the death of the organisms, their calcareous remains may accumulate on the sea floor. Provided they are not greatly diluted by weathered sediment brought into the sea from adjacent land areas, they may become cemented by chemically precipitated calcite, in the form of a lime mud, to form **limestone**. Many limestones are formed on the continental shelves at tropical and equatorial latitudes, where marine organisms thrive in the warm, sunlit waters and there is a low input of continental sediment. Sample S7 in the Kit is an example of a limestone formed in this way.

Some calcareous sediments form in the deeper ocean basins from the accumulation of calcareous microfossils that are far too small to be seen with the naked eye. These unconsolidated sediments are called calcareous **oozes**. When compacted and cemented they form fine-grained chalky limestones (Figure 17a). Well known examples of these are the White Cliffs of Dover and the North and South Downs of south-east England. However, very few deep-sea limestones are preserved in the continental geological record because they are either destroyed at subduction zones or strongly metamorphosed during inter-continental collisions.

Once exposed at the Earth's surface, limestones are very susceptible to chemical weathering because they dissolve easily in rainwater that has been acidified by atmospheric carbon dioxide. The chemical reaction that takes place is:



The weathering of limestones is a major source of both calcium and bicarbonate ions in river waters and groundwaters, which explains why water in limestone regions is 'hard': calcium carbonate is re-deposited when the water is boiled.

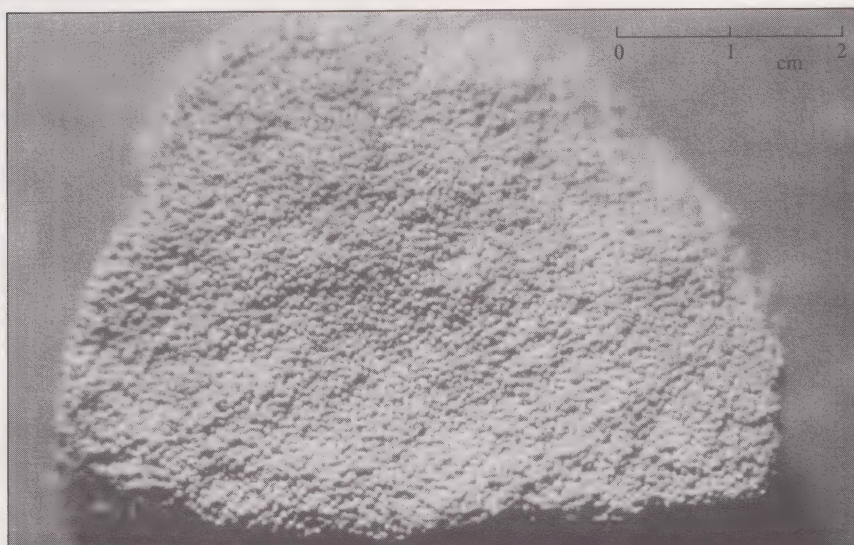
(ii) *Direct chemical precipitation*

Sediments formed from soluble cations by direct precipitation are comparatively rare, but some are of great economic importance. The best known are salt deposits, or **evaporites**, which form in isolated inland seas or in narrow, newly formed ocean basins where circulation with the other masses of seawater does not take place. Here, the evaporation of the water may exceed the input by rivers, so that the water becomes *saturated* with respect to ions, such as Na^+ , K^+ , Mg^{2+} , Ca^{2+} , Cl^- , $(\text{SO}_4)^{2-}$ and $(\text{HCO}_3)^-$, causing salts of these, such as halite (NaCl), gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) and, of course, calcite (CaCO_3), to crystallize out. Evaporites also form along arid shorelines like the present-day Persian Gulf. Here they crystallize out within the algal colonies of the mud flats. High evaporation rates are common in arid climates; thus, evaporites in rock sequences are a good indicator of ancient arid climates. Evaporites, such as the salt deposits of Cheshire, along with the wind-blown 'desert' sandstones of, for example, the Vale of Eden in Cumbria, are common in the geological record of Britain about 250 million years ago. This indicates that this part of Britain was much closer to the equator at that time.

A few limestones are precipitated directly from seawater but they are relatively rare. They require warm shallow waters (CaCO_3 is less soluble in warm water than cold water) that are well agitated by waves or currents, and particles such as sand grains or shell fragments to act as nuclei for the precipitated CaCO_3 to grow around. The resulting calcareous grains are slightly ovoid and are termed *ooliths* (Figure 17b).



(a)



(b)

FIGURE 17 Forms of limestone. (a) Calcareous microfossils, taken using a scanning electron microscope. Some rocks, such as the chalk in the White Cliffs of Dover, are made up entirely of microfossils like these. (b) Oololiths.

4.8 'FROM SNOWDON TO THE SEA' (TV PROGRAMME)

As we said in the Introduction, TV 27 'From Snowdon to the sea' may be watched at any time during your study of this Unit. However, it is particularly relevant to Sections 4 and 5.5.

The programme demonstrates the aspects of the rock cycle that are dominated by surface processes. This is done by tracing the course of the Afon (River) Glaslyn in North Wales from its source near Snowdon to its estuary at Porthmadog (Portmadoc). Plate 15 shows the Afon Glaslyn and a few of its tributaries, together with numbered locations, which are visited during the programme. The locations are:

- 1 A steep rocky stream bed that is just below the outlet from the small glacial lake in which the Afon Glaslyn rises.
- 2 A marshy area marking the spring-fed source of one of the tributaries of the main river.
- 3 A vantage point overlooking both banks of the river, giving excellent views of the upper and middle reaches of the river and its tributaries.
- 4 The head of one of the lakes into which the river flows, depositing sediments with some degree of sorting.
- 5 The river below the lake outlet, where its bed is once more a varied jumble of angular blocks of rock.
- 6 The river below the confluence of a major tributary at Beddgelert, where different flow regimes are clearly seen, and different energies of the environment can be related to different degrees of sorting of the sediments. A sieved sample of sediment from this location gave the histogram in Figure 18a.
- 7 The flat flood plain of the river in its lower reaches, where the flow is more uniform and the sediments are quite well sorted, consisting predominantly of quartz. A sieved sample from this location gave the histogram in Figure 18b. The landscape here is dominated by the processes of deposition, and stratified sediments can be seen in the river bank.
- 8 A small bay near the seaward end of the main channel provides a sufficiently low-energy environment for the suspended load of the river to be deposited as glutinous mud, consisting of clay mineral particles.

An overall view of the Portmadoc estuary shows that a considerable volume of sediment has accumulated there. Much of it is derived by weathering and erosion from Snowdonia, which offers spectacular examples of glacial erosion: amphitheatres overlooking glacial lakes, U-shaped valleys, sheer rock faces, and steep ridges. Of course, the glaciers retreated after the last ice age, but mechanical weathering still predominates in this mountainous region, and the steeper rock faces have large masses of scree at the base. The landscape is predominantly one of erosion, with weathered material being continually moved downwards into streams and river valleys, and transported on towards the sea.

The Afon Glaslyn rises in a small lake within part of the valley that has been deepened by glacial scouring. Its upper reaches are characteristic of any mountain stream: a steep rocky bed with a jumble of varied angular blocks (locality 1, Plate 15), which will become rounded and reduced in size as they are transported down the river.

The rocks of Snowdonia are ancient sedimentary and volcanic rocks that were weakly metamorphosed about 400 million years ago. This metamorphism has imparted to the rocks a slabby structure (a slaty cleavage), which gives rise to tabular blocks when the rocks are weathered. Where the slaty cleavage is particularly well developed the rocks can be called true slates; indeed, Snowdonia has long been famous for its slate industry. You will learn more about metamorphism in Section 5.4.

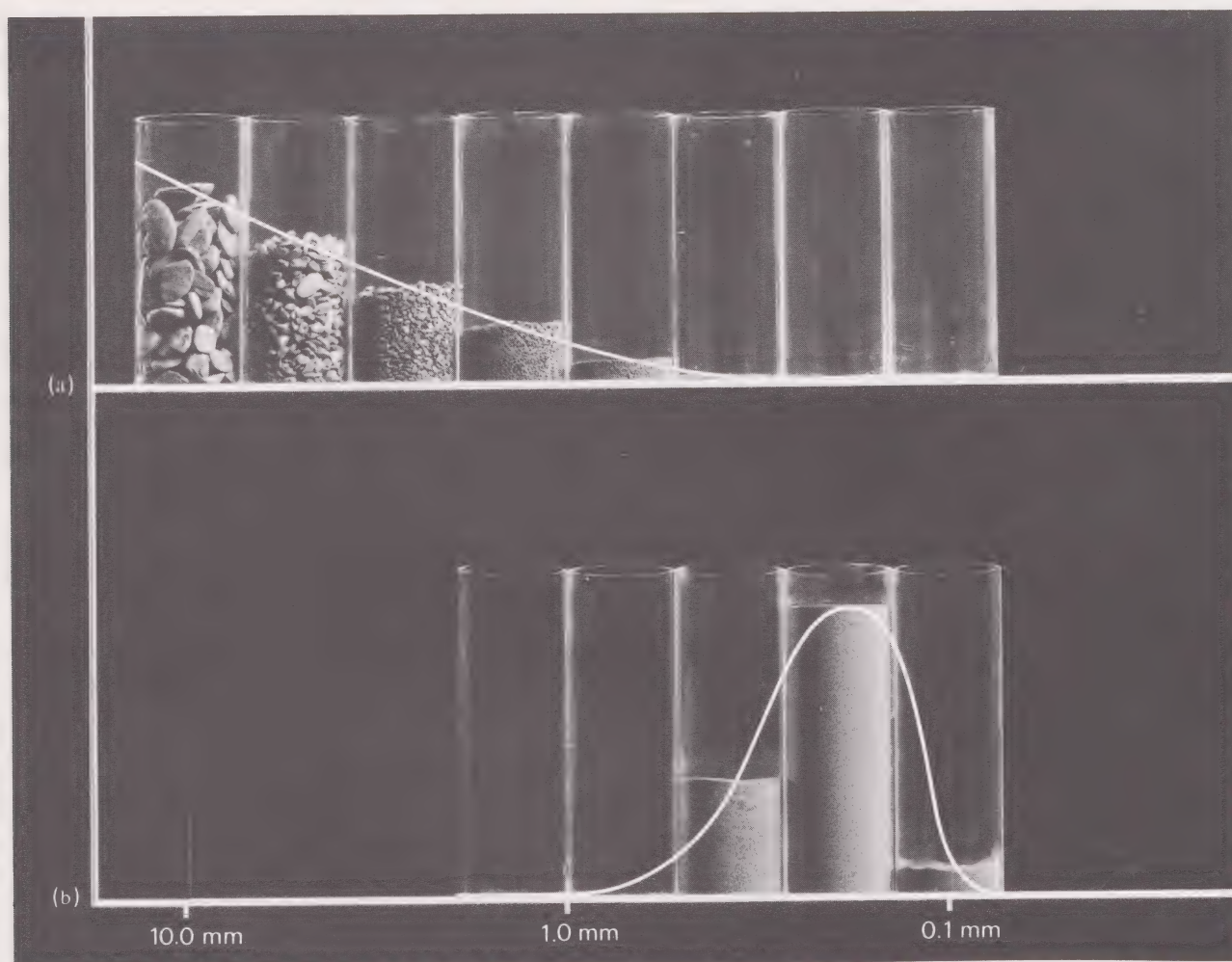


FIGURE 18 (a) Histogram showing distribution of grain sizes in sediments at locality 6 on Plate 15. In the middle reaches of the river, the sediments are not well sorted because the flow fluctuates and so the energy of the environment is rather variable. The sediments consist mainly of rock fragments. (b) Histogram showing grain size distribution of sediments at locality 7 on Plate 15. In the lower reaches of the river, the sediments are quite well sorted (compare with (a)) because the flow fluctuates less and the energy of the environment is fairly stable. The sediments consist mainly of quartz grains.

White veins of quartz can be seen in several of the rock faces, and there are blocks of quartz in the stream bed. Quartz is also an important constituent mineral of the metamorphic rocks.

The stream flows into a second small 'glacial' lake only a short distance below its source, and here its speed drops so that sediment is deposited, building a small delta out into the lake.

Other tributaries of the Glaslyn rise in springs, as a result of water percolating down into cracks and fissures in the rocks and then seeping up out of the ground (locality 2).

The river widens in its middle reaches, as more tributaries join it and the volume of water increases. Looking downstream from locality 3, we can see the river flowing through a flat meadowed area where a glacial lake is silting up owing to the deposition of fine-grained sediments from the river.

In the middle reaches of a river, fluctuations in the flow are somewhat less extreme than in the upper reaches. The fragments being transported are smaller and more rounded, having now travelled a considerable distance. In this river, the pebbles tend to be flat in shape because of the slabby nature of the source rocks; and in locality 4, too, where the river enters the lake, some sorting has taken place. Shingle characterizes the bed where the flow is rapid, while finer sands are deposited from the bedload where the river slows. In the lake itself, the water is practically still (i.e. the energy of the environment is low), and here the fine suspended material will be deposited. Lakes are very efficient sediment traps, and so where the river leaves the lake (locality 5), it is carrying a new load of sediments and has to 'start again'.

Around locality 6, the contrast in flow regimes between rapid and deeper stretches on the one hand, and between the outside and inside of bends on the other, provides an opportunity to relate the energy of environment to the degree of sorting of sediments. Rapid and turbulent flow keeps small particles moving and leaves only the coarser material. Fine material accumulates only where the flow is slower and less turbulent. There are still appreciable fluctuations in river flow in the middle reaches, and so sediment tends to be rather poorly sorted (Figure 18a). Very little suspended clay and silt-sized material is found in these sediments, because the energy of the environment is too great and the smallest particles are swept downstream.

The sands in the bedload that are sorted out from the higher-energy environment of the middle reaches begin to accumulate in the lower reaches, where they contribute to the low-gradient flood plain across which the river meanders on its way to the sea (locality 7). The sands are now quite well sorted (Figure 18b) and consist mainly of quartz, unlike the sediments from the middle reaches (localities 5 and 6), which consist mainly of rock fragments. Weathering and erosion ultimately break rocks down to their constituent mineral grains. As you know from Section 4.2, quartz is the most resistant of those minerals, which is why it forms the major component of most sands.

The flood plain (and the estuary) are flat-lying regions of sediment accumulation, and the banks of the river show evidence of stratification as a result of successive layers of flood deposits that have built up over the years.

The sands and muds that are brought down by rivers in the bedload and suspended load are ultimately deposited in the estuary, along with sediments brought in by other rivers and by waves and currents as a result of coastal erosion. The inter-tidal sand flats of an estuary are relatively high-energy regions, and very little suspended clay and silt-sized material is deposited on them. The muds accumulate in sheltered bays and inlets, where wave and current action is minimal (locality 8). The muds consist of clay mineral particles, the finest fraction derived from the weathering and erosion of rocks.

The present-day streams and rivers of Snowdonia have been in existence only since the last glacier melted, about 20 000 years ago—a very small proportion of geological time indeed. Sediments accumulating today are, however, the stuff of which tomorrow's sedimentary rocks will be made. So, by studying present-day surface processes we can interpret the record in ancient sedimentary rocks and deduce something of the conditions that prevailed in different places and at different times in the geological past. In other words, we can apply the principle of uniformitarianism—that the present is the key to the past. This is a major theme of Units 28-29.

SUMMARY OF SECTION 4

In this Section we have given you a broad introduction to the processes involved in sedimentary rock formation.

1 Most igneous and metamorphic rocks at the surface of the Earth are no longer in equilibrium with the pressure and temperature conditions of their formation, and so they undergo breakdown by physical and chemical weathering. Silicate minerals that crystallize at high temperatures, such as olivine, are more easily broken down than low-temperature minerals, such as quartz. This characteristic is related to the increasing strength of the weakest bonds in their atomic structures.

2 Chemical weathering of rocks produces soluble ions and new minerals (such as clay minerals) and leaves resistant minerals (principally quartz). The amount of chemical weathering increases with temperature and rainfall.

3 Physical weathering relies on processes such as frost shattering and the movement of material under gravity. It is, therefore, greatest in high latitudes and altitudes, where there are frequent temperature fluctuations around the freezing point of water.

4 Solid weathered material, or sediment, can be moved by water, wind or ice. In water, material moved along close to the ground (e.g. along a river

bed) is known as bedload, while that carried within the fluid medium (water, air) is called suspended load.

5 Changes in water or air current speeds cause particles to be deposited or picked up. A higher speed is necessary to erode a particle than that at which it can stay in the suspended load (fine particles) or bedload (coarser particles) before being deposited.

6 Long transportation distances and a fluid flow that is constant result in particles being sorted into different sizes. The greater the degree of selection, the better the degree of sorting.

7 The forms of sedimentary structures indicate the conditions of deposition. For example, fine laminae are characteristic of suspension deposits, whereas ripple marks and cross-bedding indicate higher water speeds involving the movement of coarse particles in the bedload.

8 Wind can transport only fine particles in suspension, but it can move coarse particles along the ground. Wind-blown sediments are usually well sorted because after a storm the wind speed gradually reduces leading to the deposition of sediments in decreasing order of size. In contrast, ice can transport a wide size range of particles many kilometres. Glacial deposits are poorly sorted, because melting ice deposits material of all sizes together.

9 The soluble products of weathering are carried to the sea, where calcium and bicarbonate now may be removed from solution by marine organisms. These and other ions may also be deposited by direct precipitation, which balances the input from rivers. Removal by marine organisms leads to the formation of fossiliferous limestones; direct precipitation forms either muddy limestones or evaporites, which include minerals, such as halite (NaCl) and gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$).

SAQ 10 Attempt this SAQ after viewing TV 27. At locality 6 in Plate 15, quartz pebbles occur in the stream bed. Why is it most unlikely that these quartz pebbles have come from the quartz veins in the rock faces in the region of Glaslyn lake (locality 1)?

SAQ 11 Which of the Kit specimens S1 (granite) and S5 (gabbro) would you expect to decompose more rapidly as the result of chemical weathering? Give a brief explanation of your answer.

SAQ 12 (a) Suggest why weathered basalt outcrops usually have a thin, surface layer of brown material.

(b) What would you expect to find as the *residual* products of a weathered granite?

(c) What would you expect to find as the *residual* products of a weathered gabbro?

SAQ 13 A river, which is flowing through a narrow gorge cut in rock, has a speed of 0.3 m s^{-1} and carries a range of particles from clay size upwards. It emerges from the gorge into a region of flat, low-lying land, and its speed is reduced suddenly to 0.02 m s^{-1} .

(a) What range of particle sizes would you expect to find deposited in the river bed where the river emerges from the rock cutting?

(b) Would you expect the sediment to be well sorted, moderately sorted, or poorly sorted?

SAQ 14 Examine Kit specimen S6 (sandstone) closely with your hand lens.

(a) Suggest in relative terms what the energy of the transporting medium might have been.

(b) Suggest in relative terms for how long the material underwent transportation.

TABLE 13

Ion	Concentration in river water	Concentration in seawater
Ca^{2+}	13	1.2
Na^+	1.7	31
HCO_3^-	49	0.40
$(\text{SiO}_4)^{4-}$	11	negligible

SAQ 15 In Table 13 are shown the concentrations (as percentages of the total dissolved solids) of various ions in seawater and river water. (The river water data have been corrected for the salts recycled by sea spray.) Why are there such large differences between the two columns?

METAMORPHIC ROCKS

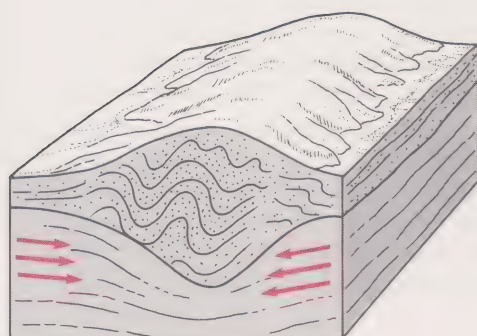
METAMORPHISM

BRITTLE DEFORMATION OF ROCKS

PLASTIC DEFORMATION OF ROCKS

CREEP IN ROCKS

JOINTS



5 TECTONIC AND METAMORPHIC PROCESSES

In the previous Sections we examined two of the essential stages in the rock cycle: the formation of igneous and sedimentary rocks. In this concluding Section, we follow the rock cycle a stage further, and go on to look at the transformation processes that affect rocks that are subjected to changes in physical conditions. In particular, we look at what happens when rocks are subjected to mechanical forces—what happens when they are squeezed and distorted—and also what happens when they are subjected to extreme physical conditions of temperature and pressure. The two processes together give rise to a whole new group of rocks, called the **metamorphic rocks**. The group includes all those rocks—of either sedimentary or igneous origin—that have been substantially changed since their formation. The important point to remember about this changing, or **metamorphism**, of existing rocks is that all the changes take place in the *solid state* (i.e. without melting). The transition from metamorphic to igneous conditions is marked by the melting point of the rock; below the melting temperature any changes will produce metamorphic rock. The changes may be barely perceptible, or they may be extreme, so that entirely new rocks are produced.

Both tectonic and metamorphic processes are extremely important to geologists, because they determine the final appearance of a rock as it is seen in the field—it is quite unusual to see a rock that shows no sign of *either* process. In terms of the rock cycle, we have reached the stage illustrated in the third part of Figure 2.

5.1 TECTONIC PROCESSES

Tectonic processes are those concerned with the mechanical deformation of rocks. Almost all rocks are deformed in some way, although this may not always be obvious. As soon as one layer of sediment is deposited on top of another, the lower layer is squashed and compacted. Sometimes it is reduced by as much as 80% of its original thickness. When sedimentary rocks are subjected to sideways compression they may crumple up much more conspicuously by forming folds. The word *tectonic* is derived from the Greek word for *builder*; it covers those mechanical processes by which rocks are built up into the complexities we eventually see at the surface.

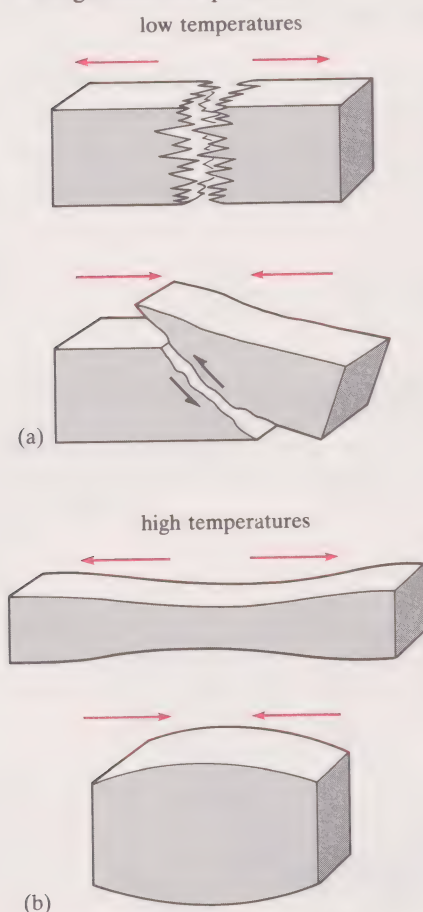
You should recall from Units 5–6 that when a given stress is applied to a rock, it responds by deforming, and the deformation is measured in terms of *strain*. Several factors control the way in which the rock will be deformed.

□ What do you think they are?

■ The most obvious one, of course, is the nature of the rock itself, which is measured in terms of the two elastic moduli that you encountered in Units 5–6. Less obvious, but more important for our present purposes, are the physical conditions in which the process takes place. The *temperature* of the rock is important as are stress, whether it is *compression* or *extension*, and the *time* over which the stress is applied.

It is common sense to think that if a rock specimen is subjected to an extension, it will eventually break cleanly apart. Under compression, the rock may be ruptured, though not with a clean, smooth fracture (Figure 19a). When rocks respond to deformation by snapping or rupturing, this behaviour is known, not surprisingly, as **brittle deformation**. At low temperatures, rocks usually behave in this way. At higher temperatures, however, a rock specimen will behave rather differently. Under tension, it will tend to stretch and become thinner rather than snapping apart,

FIGURE 19 (a) At low temperatures a rock specimen subjected to tension will snap; under compression, it will rupture along an inclined failure surface. (b) At high temperatures, the same specimens are more likely to stretch under tension, or bulge under compression.



whereas under compression it will not rupture, but will merely bulge outwards like a beer barrel (Figure 19b). This is called **plastic deformation**, and a temperature of over 100°C is usually required to bring it about. The higher the temperature, the more easily the rock is deformed plastically.

When a rock is stressed, it may respond with an *immediate* plastic or brittle deformation. For brittle deformation, there is a critical stress which must be reached prior to rupture. If the stress is maintained continuously over a long period, and if it is less than the critical stress, the rock may continue to deform indefinitely. On a small scale, this phenomenon is known as **creep**. Like plastic deformation, creep is accelerated at high temperatures. Avoidance of creep is important to engineers designing large mechanical structures intended for a long life, and it causes them considerable headaches. But creep is even more important in geology because geological timescales are much longer than those with which engineers normally deal. Almost all of the rock structures that you are likely to see are the result of creep deformation.

5.2 STRUCTURES PRODUCED UNDER TENSION

Structures produced under tension are perhaps the simplest and most widespread rock structures. When an igneous rock cools, it tends to shrink slightly, and this may set up tensional stresses throughout a volume of rock whose boundaries are already defined by crystallization. A rock often responds by breaking up into a well-defined pattern of polygonal fractures, known as **joints**, which give the rock a highly distinctive appearance. The Giant's Causeway in Antrim (Plate 3) is a magnificent example of this phenomenon. Joints are simply cracks in the rock, where it has split or fractured *without* any movement taking place *along* the crack. They are extremely common in all kinds of rock, and can be seen in almost any outcrop of rock. Figure 20 shows joints in a sedimentary rock, a limestone.



FIGURE 20 Joints in limestone. This photograph was taken in Dorset and shows where the limestone is being quarried for building stone, taking advantage of the joint pattern within the rock.

FAULT

NORMAL FAULT

THRUST

REVERSE FAULT

FOLDS

Terms in AV sequence:

METAMORPHIC GRADE

CLEAVAGE (IN ROCKS)

ROCK TEXTURE

Qualitatively different from joints are **faults**. These are planar cracks running through rocks but, unlike joints, movement takes place along the fracture plane, causing a relative offset between features on different sides of the fault. You were introduced to fault movement in relation to earthquakes in the earlier Earth sciences Units. By far the most common faults are called **normal faults**, where the rock on one side of the fault has simply dropped down relative to that on the other side (Figure 21).

- ☐ Do you suppose that faults would be formed in rocks under tension at high temperatures? If not, what do you suppose would happen?
- ☒ With increasing temperature, rocks would be less likely to break and form faults, and more likely to stretch out plastically, like heated pitch.

5.3 STRUCTURES PRODUCED UNDER COMPRESSION

When rocks are squeezed hard, especially at low temperatures, they rupture. The commonest kinds of structure produced in this way are known as **thrusts** (Figure 22c). A thrust is produced when a large mass of rock breaks in two along a low-angled plane, and the upper block overrides the

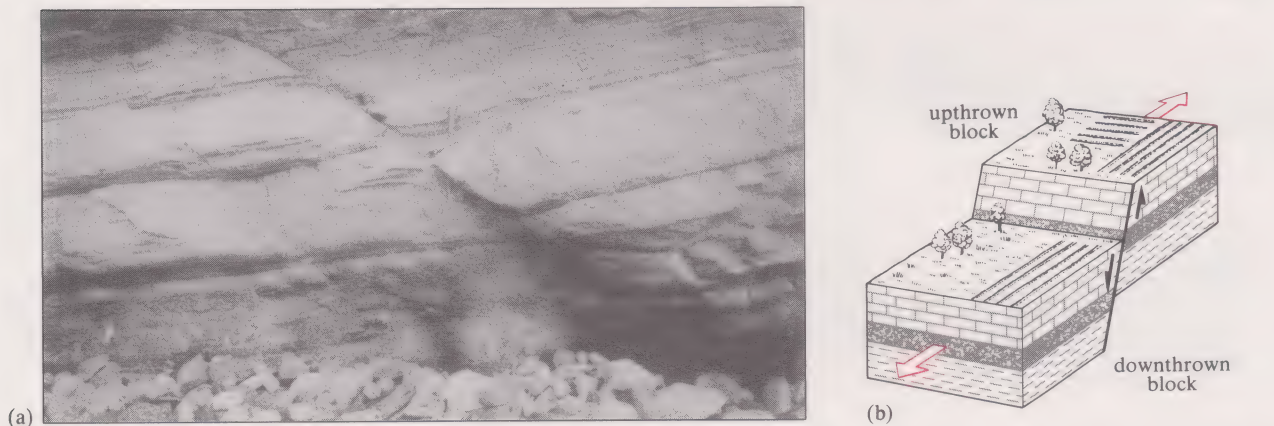


FIGURE 21 (a) A typical small-scale normal fault in layered sedimentary rock. The fault is in the centre of the photograph and runs from top left to bottom right. The rocks on the right hand side of the fault have moved down relative to those on the left of the fault, by a distance of about the length of the geological hammer, say 30–50 cm. (b) This is a block model showing the formation of a normal fault. Compare this with the fault in (a).

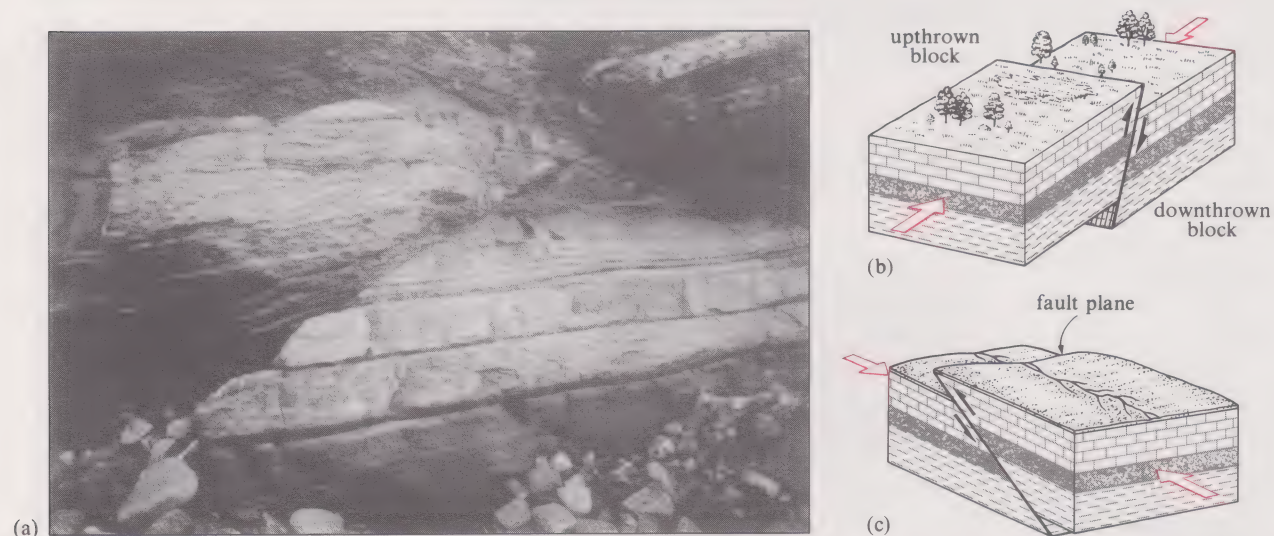


FIGURE 22 (a) A typical small-scale reverse fault, which you should compare with Figure 21a. The rocks on the left-hand side of the fault have moved *up* over those on the right of the fault. (b) Block model showing the formation of a reverse fault. Compare this with the fault shown in (a). (c) Block model of a thrust fault, where the angle that the fault plane makes with the horizontal is much lower than for the reverse fault in (a) and (b).

lower one. Some of these moving blocks may be huge (of the order of kilometres), and cases are known of thrust blocks moving for tens and even hundreds of kilometres. On a smaller scale, high-angle compressional structures are known as **reverse faults** (Figure 22a and b).

When rocks are subjected to compression at higher temperatures over long periods, they may not fracture but **fold** instead. This is particularly evident when layered sedimentary rocks are involved. The overall shortening of the layers absorbs the compressive stress. Many kilometres of shortening can be seen in mountain chains such as the Alps and, on a smaller, but still impressive scale in many British sea-cliffs and crags (Figure 23).



FIGURE 23 Folding in the cliffs of Northumberland. Here, students at a Summer School associated with a third-level Earth sciences course are examining one of a series of folds in sedimentary rocks.

5.4 METAMORPHIC PROCESSES

Before reading any further in this Section, you should work through the tape sequence 'Metamorphic rocks', which is on Tape 4, Side 2, Band 1. This will ensure that you are familiar with the rocks and concepts discussed in this Section. You will need the rock samples and hand lens. You will also need to refer to Plates 16 to 20 at the back of this binding. During the sequence you should complete Table 14.

TABLE 14 Characteristics of metamorphic rock specimens in the Kit.

	Texture	Minerals present	Conditions of formation (pressure and temperature)
slate	fine-grained, crystalline	muscovite (recrystallized clay minerals)	low temperature (below 300 °C); low pressure
S8 phyllite		muscovite (recrystallized clay minerals)	
S9 schist			
S10 gneiss		dark: light:	

CONTACT METAMORPHISM

REGIONAL METAMORPHISM

THERMAL GRADIENT

SLATE

PHYLLITE

SCHIST

GNEISS

Metamorphic rocks are, as you should now be aware, rocks that have been derived from other rocks through chemical and physical processes. (The word *metamorphism* has Greek roots, which mean something like ‘changing shape’, but it is used to cover any change induced in a solid rock due to variations in pressure or temperature.) There is nothing very mysterious about metamorphic processes in rocks, although the details of the processes may be complex. As you heard on the tape sequence, an everyday example of metamorphism is the firing of clay to make pottery; another is the baking of cakes.

The simplest example of metamorphism in geology is **contact metamorphism**, which describes the changes brought about in a rock when it is intruded by a body of hot magma. You will recall that when plutonic rocks are intruded into other rocks, they chill against their surroundings, and the intrusive body develops a fine-grained chilled margin. The surrounding rocks do not escape unchanged, and may be quite severely affected by the pluton.

- ☐ What factors are likely to determine the changes in the rock?
- ☒ The temperature of the pluton and, less obviously, its size. The susceptibility of the surrounding rocks to metamorphism is also significant (as you will see later).

The size of the intrusion is important because rocks are such poor conductors of heat that the volume heated by a small body, such as a one-metre wide dyke, is almost negligible, even though the magma in the dyke may be a thousand degrees hotter than its surroundings. By contrast, a one-kilometre-thick pluton will cool slowly enough to heat up a large volume of rock. It follows that the nearer they are to the pluton, the hotter the surrounding rocks get.

Contact metamorphism has two extremely characteristic effects. First, the surrounding rocks get ‘baked’—almost literally. They become extremely hard (like an over-cooked cake) and very splintery. Quarrymen of old had a special name for these hard rocks; they called them *hornfels*. Second, and more important for our purposes, rocks affected by contact metamorphism often have a spotted appearance (Figure 24). This is due to the growth of *new minerals* within them. The new minerals grow in response to the new physical conditions introduced by the intrusion of the pluton, and the nature of the minerals that grow is directly controlled by the temperature reached in the contact zone around the intrusion.

Contact metamorphism offers an elegant illustration of metamorphic processes—it is easy to see why the rocks were heated, and to trace the variations in the metamorphosed rock around the pluton. Observations of these changes reveal an important fact: the *chemical composition* of the metamorphosed rocks is not affected. The changes that take place are purely *mineralogical*.

An analogy can be drawn with a cake baked in an oven: the finished cake looks nothing like the initial mixture that went into the oven, but *nothing has been added to it*. And the only material that has been subtracted is water in the form of steam, and perhaps a little carbon dioxide.

Furthermore, just as the final flavour and texture of the cake depend on the conditions within the oven—especially the temperature—so the minerals that grow in metamorphic processes are controlled by the prevailing conditions of temperature and pressure. *Different minerals are characteristic of different sets of physical conditions.*

This key fact enables us to make sense of the most important kind of metamorphism, **regional metamorphism**. As its name implies, this process affects huge areas of rock, often extending over tens of thousands of square kilometres. In fact, most of the rocks making up the shallowest parts of the continental crust of the Earth are regionally metamorphosed rocks.

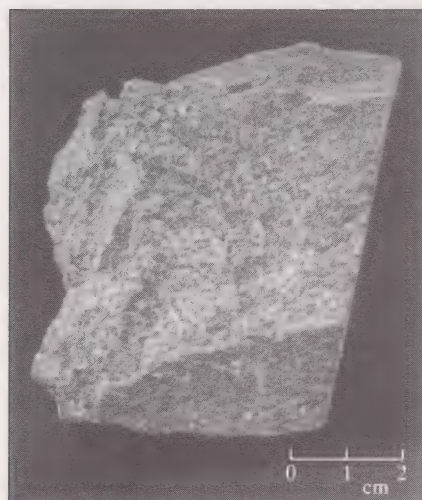


FIGURE 24 Spotty contact-metamorphosed slaty rock. The spots on this rock were formed by metamorphism near the contact of a granite in North Wales.

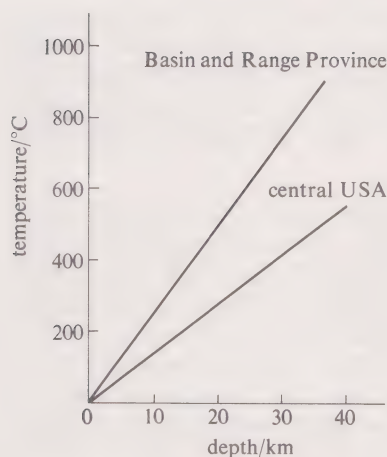


FIGURE 25 Variation of temperature with depth for two areas within the USA. The Basin and Range Province is in the foothills of the Rockies.

- ☐ Where do you think such large masses of rock could be subjected to high temperatures and pressures?
- ☒ You may recall from Units 7–8 that destructive plate margins, where mountain belts are born, are also characterized by metamorphic processes.

Both constructive and destructive plate margins are characterized by high *heat flow*. In both of these environments, therefore, one would expect to find deeply buried rocks that have been heated to unusually high temperatures for long periods. We shall concern ourselves here only with what goes on at destructive plate margins.

The rate at which temperature increases with depth in a given region is measured in degrees celsius per kilometre. This is known as the **thermal gradient**. Figure 25 shows the variation of temperature with depth for two quite different areas in the USA: one on the edge of the Rockies, and the other right in the middle of the continent.

- ☐ What are the thermal gradients in each case, and what is the temperature at a depth of 20 km in each case?
- ☒ The thermal gradient beneath the stable, continental part of the USA is about $13^{\circ}\text{C km}^{-1}$, but it is $25^{\circ}\text{C km}^{-1}$ beneath the mountain belt. Therefore, the temperature at a depth of 20 km is 500°C beneath the Rockies, which is nearly twice that beneath the central part of the USA.

It's clear, then, that one does not have to go far down beneath the Rockies to encounter high temperatures. So what happens to the rocks down there? Well, we cannot burrow down that far beneath the Rockies to see for ourselves, but we *can* look at the roots of older mountain belts that have been brought back to the surface by uplift and erosion. In areas such as the Highlands of Scotland, where rocks involved in mountain building events over 400 million years ago are exposed, we can trace exactly what happens to rocks that start life as ordinary, wet sediments—the kind of thing you might find forming today on a beach or in a river.

Consider mud. When fresh, mud has properties that scarcely need description—it is wet, soft and highly compressible. When deposited in substantial thickness in, say, an estuarine environment, muds experience a number of changes as water is squeezed out of them and they become hardened. Eventually, the mud becomes cemented and compacted, and a mudrock (sometimes called mudstone) results. These changes, however, do not warrant the term *metamorphism*. This is reserved for the changes that take place in response to heat and pressure. If subjected to a mild degree of metamorphism, the mud will grow new minerals, usually flaky, micaceous ones, and will acquire a perceptible lamination, producing a **slate** with a well-defined cleavage. With increase in pressure the slate itself will become deformed and a **phyllite**, like your specimen S8, will be produced. The phyllite started life as a muddy sediment, but sufficient new mineral growth has taken place within it to give the rock a distinctive sheen, and an excellent lamination. Such mild regional metamorphism, where temperatures are lower than about 500°C , is known as *low-grade* metamorphism.

A **schist** (specimen S9) is the next stage in the process. It is much more coarsely crystalline, and it is difficult to see much of its original nature. Although it may once have been a sediment similar to that from which the phyllite formed, all trace of sedimentary properties has gone. The **gneiss** (S10) is the last step in the chain and at this stage the rock is very close to becoming granite. As you can see from your specimens, the texture of the gneiss is much more like that of the granite than the phyllite; the only indication of its metamorphic nature is the well-displayed planar banding. While slates, phyllites and schists are all usually generated from mudrocks, you should note that gneisses may arise from a wide variety of pre-existing rocks. Granites and gabbros, for instance, can easily be converted to granitic or gabbroic gneisses if the temperature and pressure are high enough. Gneisses are formed at temperatures greater than 600°C , under what are known as high-grade metamorphic conditions.

In this Section, we have been dealing with metamorphosed rocks that are formed at increasing metamorphic grades. You should now begin to see a link emerging between the most highly metamorphosed rocks and igneous rocks.

- ☐ Refer back to Figure 25 and look once more at the temperature–depth profile beneath the Rockies. Assuming that the temperature continues to increase with depth at the rate indicated, what will be the temperature at a depth of 40 kilometres?

■ It will be about 1 000 °C.

The importance of this is that, because granites have low melting points (around 700–850 °C, much lower than basalts), one might expect to find granite magmas at depths of between 30 and 40 kilometres below the Rockies. So you can see that in the most extreme conditions of regional metamorphism, sedimentary rocks may be heated sufficiently to melt them and for granitic magmas to be produced.

It is important to emphasize that wholesale, complete melting of metamorphosed sedimentary rocks would *not* take place. Just as *partial* melting of mantle peridotite forms basaltic magmas (Section 3.3), and partial melting of andesite yields granitic magmas (Section 3.5), so partial melting of the most intensely metamorphosed crustal rocks may produce granitic magmas. In general, only a few per cent of the total is melted, and this, of course, is the granitic component, which may rise through the crust as magma. The other components, which have higher melting points, remain in a solid state.

In the deep roots of mountain chains, then, igneous and metamorphic processes converge. This is particularly true of the sites of colliding continents, where granitic magmas can be unequivocally related to the partial melting of metamorphosed sedimentary rocks. In general, however, large volumes of granite are not produced simply by melting sediments. It is thought that most granitic magmas arise at destructive plate margins where andesites, produced by partial melting of subducted ocean crust, undergo successive stages of ‘refining’ to generate more siliceous melts as described in Section 3. It is thought that a small proportion of material from the lower continental crust at Andean margins may also melt and contribute to the granitic magma from this environment. The magmas that are produced move upwards and away from their source in the lower crust, and contribute to the large granite batholiths formed at Andean plate margins.

Thus the rock cycle is completed: igneous rocks are eroded to form sediments, the sediments are metamorphosed and may eventually be melted to contribute towards a fresh generation of igneous rocks (Figure 26). This cycle is extremely important, so it is worthwhile summarizing all the steps that may occur between the mantle peridotite and the granites formed in continental crust:

- 1 Partial melting of mantle peridotite at constructive plate margins produces basaltic magmas, which form oceanic crust.
- 2 Basaltic oceanic crust subducted at destructive plate margins is partially melted, along with mantle peridotite that overlies the subducted slab, and produces andesitic magmas that form the volcanoes of island arcs.
- 3 Sustained igneous activity, particularly in a continental setting where great thicknesses of andesitic material are present, produces more siliceous magmas, including granites.
- 4 Andesitic and granitic igneous rocks are exposed at the surface and are then eroded to form fresh sediments.
- 5 These sediments are converted into sedimentary rocks that are deeply buried and subjected to regional metamorphism at destructive plate margins or at the sites of colliding continents.
- 6 Regional metamorphism produces progressive changes in sediments: from sediment, through slate and phyllite to schist and, perhaps, to gneiss.

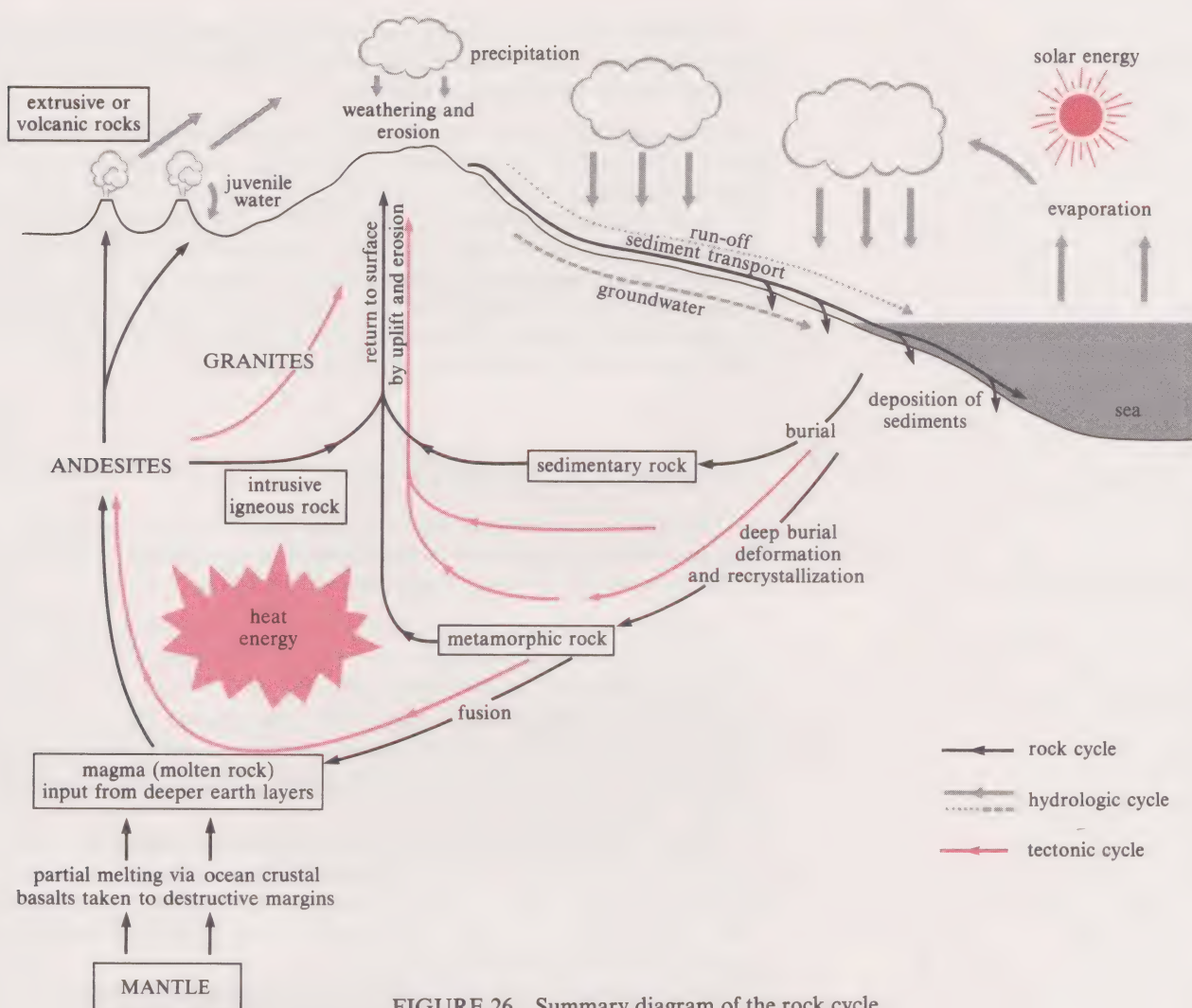


FIGURE 26 Summary diagram of the rock cycle.

7 Eventually, partial melting of schists and gneisses produces small volumes of granitic magmas, which are incorporated into the spectrum of magmas associated with igneous activity above subduction zones. Similarly crustally-derived granites form in the collision zones between converging continents.

8 The new granitic magmas move away from the melt zone and upwards in the crust to become igneous intrusions.

9 Cover rocks are eroded, igneous intrusions are exposed and then these are eroded to form a fresh generation of sediments.

5.5 THE ROCK CYCLE—A CONCLUDING COMMENT

A unifying theme in this Unit has been the rock cycle: the ceaseless erosion of rocks, the transport of sediment by rivers, and the deposition of sediment to form new rocks. Parts of this cycle are so well known that they have become entrenched in popular literature. Travel brochures use advertisements such as 'brooding mountains sculpted by nature through long aeons of ceaseless toil'.

The extreme consistency of the geological processes involved in the rock cycle led, in the 19th century, to the formulation of a key concept in geology, that of uniformitarianism, which you will learn about in Units 28–29. Proponents of this concept argued that the processes we see operating on Earth today are the same as those that have always operated, and

that we can understand old rocks by interpreting them in terms of processes operating today. This concept holds up extremely well in relation to the rock cycle, with one important exception.

The exception lies in the formation of continental crust from mantle peridotite. The rock cycle, as we know it, simply could not get under way without the formation of the continents. These, as we have seen, are formed by the irreversible addition of andesites by a two-stage partial melting process, at constructive and destructive margins, from mantle peridotite. Although some continental sediments are transported into the ocean and become subducted it seems that, on balance, there is a slow and steady increase in the amount of continental crust. There is, in effect, therefore, a steady input of new material into the rock cycle.

SUMMARY OF SECTION 5

- 1 The study of tectonic processes shows that at low temperatures rocks snap and fracture but at high temperatures they stretch and bend. These are known respectively as brittle and plastic deformation.
- 2 Joints are produced as a result of tension and are found in all types of rock. Faults are formed when the rocks each side of the planar crack move relative to each other. Normal faults result from tension.
- 3 Thrust and reverse faults result from compressive forces at low temperatures; one block is pushed up and over the adjacent block. At higher temperatures, rocks bend into folds under compression, or stretch to become thinner under tension.
- 4 Contact metamorphism occurs when hot magma affects the rocks that are directly in contact with it. Regional metamorphism is a larger-scale phenomenon that occurs at depth owing to the regional thermal gradient. Most continental rocks and some oceanic rocks have experienced some degree of regional metamorphism.
- 5 New minerals grow in rocks as a result of metamorphic processes; when directional pressure is involved, platy minerals in particular develop at right angles to the direction of pressure and so impart cleavage to the entire rock.
- 6 As mudrocks are metamorphosed they pass through a metamorphic sequence from slate to phyllite to schist. As they do so, the size and complexity of the metamorphic fabric and mineralogy increases.
- 7 At the highest temperatures and pressures (high-grade metamorphism), the melting point of granitic rocks is exceeded and igneous magmas are formed.

SAQ 16 Which two of the following statements associated with tectonic processes are correct?

- (a) Rocks are stronger at high temperatures than at low ones.
- (b) Rocks may break or fracture when subjected to either extensional or compressive stresses.
- (c) Creep will not take place in rocks unless a certain minimum stress is applied.
- (d) The rate at which creep takes place is independent of other factors.
- (e) Rocks are stronger in tension than they are in compression.
- (f) Joints in rocks are always the result of compressive stresses.

SAQ 17 What kinds of structure might be produced in rocks by compressive stresses at low temperatures; and what structures would be more likely at high temperatures? Give reasons for your answers.

SAQ 18 If granitic rocks melt at a temperature of 850°C , at what depth would you expect to find partial melting of crustal rocks taking place in a region where the thermal gradient is $15^{\circ}\text{C km}^{-1}$? Assume that the surface temperature is 20°C .

SAQ 19 Match each of the following rock types (1–4) with one of the sets of conditions in the key (A–D).

Rocks

- 1 Partially melted sediment
- 2 Phyllite
- 3 Hornfels
- 4 Gneiss

KEY

- A Contact metamorphism
- B High-grade regional metamorphism
- C Thermal gradient of $28^{\circ}\text{C km}^{-1}$ extrapolated to 30 km
- D Low-grade regional metamorphism

SAQ 20 Arrange the following four metamorphic rocks in increasing order of metamorphic grade.

gneiss, phyllite, schist, slate

SAQ 21 Explain how you would distinguish between metamorphic banding and sedimentary layering in rock samples. How is each formed?

SAQ 22 At the beginning of this Unit we posed four questions, which you should now be able to answer:

- (a) Why are the volcanoes of the Andes more dangerous than those of Hawaii?
- (b) How are sediments formed?
- (c) What are the processes involved in turning mud into slate?
- (d) Are the processes in (a)–(c) above linked? If so, through what mechanism or model?

Only brief answers are required!

OBJECTIVES FOR UNIT 27

After you have completed your work on this Unit, you should be able to:

- 1 Explain the meaning of, and use correctly, all the terms flagged in the text.
- 2 Recognize the characteristic physical properties, such as cleavage, lustre, density and hardness, of common minerals. (*1st AV*)
- 3 Relate the colour of a silicate mineral to its chemical composition, and the chemical composition to its structure and temperature of formation. (*SAQ 1, ITQ 2*)
- 4 Relate the chemical composition and structure of silicate minerals to their densities. (*SAQs 1 and 2*)
- 5 From a hand specimen of igneous rock, make predictions about the mineral composition, the temperature of formation and order of mineral crystallization, and the density of the rock. (*SAQs 1 and 2*)
- 6 Describe the possible sources of the Earth's internal energy. (*SAQ 3*)
- 7 Explain systematically the ways in which the compositions of igneous rocks can be changed and new rocks formed, by partial melting and fractional crystallization. Solve simple problems relating to these processes. (*SAQs 4, 7 and 8–10*)
- 8 Summarize the main differences between volcanic and plutonic rocks. (*SAQ 3*)
- 9 Explain how continents are formed by destructive plate margin processes. (*SAQs 7 and 10*)
- 10 Relate chemical weathering processes to the mineral compositions of igneous rocks. (*SAQs 12, 13 and 15*)
- 11 Relate the physical processes of sedimentation in water environments to sedimentary rocks. (*SAQs 14 and 15, ITQs 5 and 7*)

- 12 Explain the role of the soluble products of chemical weathering in the formation of sedimentary rocks. (SAQ 16)
- 13 Summarize the ways in which rocks can be deformed. (SAQs 16 and 17)
- 14 Describe and explain the reasons for the formation of joints, faults and folds. (SAQs 16 and 17)
- 15 Describe the nature of metamorphism and explain the relationship between metamorphism and thermal gradient. (SAQs 18 and 19)
- 16 Summarize the progression of different rock types produced by regional metamorphism. (SAQs 19 and 20, 2nd AV)
- 17 Distinguish between sedimentary layering and metamorphic banding in a hand specimen and explain how each is formed. (SAQ 21, 2nd AV)

ITQ ANSWERS AND COMMENTS

ITQ 1 (a) High-temperature forms are magnesium rich, whereas low-temperature forms are iron rich.

(b) High-temperature plagioclase feldspars are calcium rich, whereas low-temperature plagioclase feldspars are sodium rich.

(c) With falling temperature, Fe in silicate minerals increases at the expense of Mg, but in general Fe and Mg minerals crystallize at high temperatures. Low-temperature minerals are rich in Na, K and Si.

ITQ 2 1 g cm^{-3} .

In 1 m^3 there are $10^2 \text{ cm} \times 10^2 \text{ cm} \times 10^2 \text{ cm} = 10^6 \text{ cm}^3$.

In 1 kg there are 10^3 g , so in $1 \times 10^3 \text{ kg}$ there are 10^6 g .

So, $1 \times 10^3 \text{ kg m}^{-3} = \frac{1 \times 10^6}{10^6} \text{ g cm}^{-3} = 1 \text{ g cm}^{-3}$.

ITQ 3 (a) 10 g ; (b) 10^{-1} N .

(a) $\text{mass} = \text{volume} \times \text{density}$
 $= 10 \text{ cm}^3 \times 1 \text{ g cm}^{-3}$
 $= 10 \text{ g}$

(b) $\text{weight} = \text{mass} \times \text{acceleration due to gravity}$
 $= 10 \text{ g} \times 1000 \text{ cm s}^{-2}$
 $= 10000 \text{ g cm s}^{-2} = 10^4 \text{ g cm s}^{-2}$

Since $1 \text{ kg m s}^{-2} = 1 \text{ N}$ (Unit 3), the answer can be converted into newtons (the SI units of weight) by dividing by 10^3 (to convert g to kg), and dividing by 10^2 (to convert cm to m). So,

$\text{weight} = 10^4/10^5 \text{ kg m s}^{-2}$
 $= 10^{-1} \text{ kg m s}^{-2}$
 $= 10^{-1} \text{ N}$

ITQ 4 (a) $2.59 \times 10^3 \text{ kg m}^{-3}$.

(a) First you need to find the volume of the sample. You can find its magnitude simply by subtracting the value in row 3 from that in row 2. This gives a value of 2.97 for the volume of the sample, and the units will be cm^3 . Density is obtained by dividing the mass by the volume:

$$\begin{aligned} \text{density} &= \frac{7.68 \text{ g}}{2.97 \text{ cm}^3} \\ &= 2.59 \text{ g cm}^{-3}, \text{ or } 2.59 \times 10^3 \text{ kg m}^{-3} \end{aligned}$$

(b) The density of pure quartz is $2.65 \times 10^3 \text{ kg m}^{-3}$ and that of pyrite is around $5.00 \times 10^3 \text{ kg m}^{-3}$. You should get results quite close to these values, say $\pm 0.2 \text{ g cm}^{-3}$ (i.e. $\pm 0.2 \times 10^3 \text{ kg m}^{-3}$, or within about 8%); we tried the same thing at home, using the same equipment, and found only small differences.

(c) If you were to measure the density of the feldspar crystal you might get a slightly different result, depending upon the exact chemical composition of your sample. Typical values for the density of feldspar are $2.56\text{--}2.70 \times 10^3 \text{ kg m}^{-3}$, as shown in Figure 4. The densities of quartz and feldspar are therefore similar, but these minerals are 20–25% less dense than olivine and only about half the density of pyrite.

ITQ 5 The olivines are found only in igneous rocks, whereas clay minerals are found only in sedimentary rocks, and garnets only in metamorphic rocks. Amphiboles occur both in igneous and metamorphic rocks, but not in sedimentary ones. There are clearly more points of similarity between metamorphic and igneous rocks than between sedimentary and igneous rocks.

ITQ 6 (a) The granite and the sandstone contain most potassium; the K_2O values are 4.2% and 4.8% respectively. Peridotite contains the least (0.03%).

(b) The ratio of potassium in the sandstone to that in the peridotite is 4.8:0.03, i.e. 160 to 1.

(c) The continental crustal rocks, granite and sandstone, followed by andesite, are richest in potassium. Oceanic crustal rocks (basalts) are far less rich. Mantle rocks (peridotite) are least rich in potassium.

ITQ 7 Since pyroxene and feldspar crystallize at a lower temperature than olivine, it will start to melt earlier, and thus the first liquid will tend to have a pyroxene–feldspar composition. Similarly, since iron-rich minerals crystallize at lower temperatures than their magnesium-rich counterparts, the first liquid will tend to have more iron than magnesium.

ITQ 8 There is a progressive *increase* in SiO_2 , Na_2O and K_2O in going from basalt to granite, and a *decrease* in MgO , $\text{Fe}_2\text{O}_3 + \text{FeO}$, TiO_2 and CaO . The implication is that, starting with a liquid of basaltic composition, rocks of any intermediate composition up to granite could be made by subtracting the right elements, particularly Mg, Fe, Ti and Ca. The minerals required to make this subtraction are mainly high-temperature forms of plagioclase feldspar and pyroxene (Figure 4).

In this example, the MgO in the olivine-depleted liquid is less than that in andesite. In nature olivine would be joined by Mg-pyroxene and Ca-feldspar early in the fractionation.

ITQ 9 (a) Cations such as K^+ , Na^+ and Ca^{2+} (along with a little silica) would be leached into solution leaving a residue of clay minerals derived from the feldspar. A similar fate might befall the mica, but if it were biotite mica, then a small amount of iron oxide and hydroxide residue might occur, too. The quartz would be unaffected and so would provide a residue of quartz grains.

(b) Olivines and pyroxenes would decompose completely, releasing Mg^{2+} and silica into solution. This would leave only a residue of iron oxides and hydroxides to accompany the clay residue formed from the breakdown of feldspar (as in (a)).

(c) The calcium carbonate cement would dissolve releasing calcium and bicarbonate ions into solution and leaving behind a sandy deposit of residual quartz grains, unaffected by chemical weathering.

ITQ 10 (a) A horizontal line drawn across the graph at 0.3 m s^{-1} intersects the curve for deposition at a grain size corresponding to about 10 mm (1 cm). Grains this size would have to move as bedload.

(b) Only grains coarser than 10 mm will have stopped moving altogether, so the sediment on the bed will be coarse-grained gravel and pebbles.

(c) A vertical line drawn up the graph intersects the curve for deposition at about 0.008 m s^{-1} (less than 1 cm s^{-1}), which means that very fine grains are only deposited when water movement is almost negligible.

SAQ ANSWERS AND COMMENTS

SAQ 1 (a) Your completed Table 8 should resemble Table 15. Note that in the granite (S1) you would expect feldspars and micas to crystallize at about the same temperature; in the gabbro, pyroxenes and feldspars will crystallize at about the same temperature.

(b) The colour difference is a result of the mineral proportions. The granite (S1) contains only 10% dark ferromagnesian minerals and therefore appears pale coloured. The gabbro (S5) and peridotite (S4) contain 40% and 100% ferromagnesian minerals respectively, and so appear much darker coloured.

TABLE 15 Suggested answer to SAQ 1.

Rock sample	Average mineral composition	Colour of mineral	Overall rock colour (light or dark)	Density/ kg m^{-3}	Relative temperature of formation (high, moderate, low)	Expected sequence of mineral crystallization
S1 GRANITE	10% biotite mica	brown-black	light	$2.4\text{--}2.6 \times 10^3$	low	mica + feldspars
	60%, mainly K-feldspar	pinkish-white				
	30% quartz	colourless				quartz
S5 GABBRO	10% olivine	difficult to see any—may be occasional brown crystals	dark	3.0×10^3	moderate	olivine
	30% pyroxene	dark greeny-black				pyroxene + feldspar
	60% plagioclase feldspar	white				
S4 PERIDOTITE	85% olivine	olive green	dark	3.3×10^3	high	olivine
	15% pyroxene	dark greeny-black				pyroxene

(c) The ferromagnesian minerals, olivines and pyroxenes, contain the high-density elements and also have close-packed structures. A rock like S4, which contains all olivines and pyroxene, will therefore have a high density. A rock like S1, which contains 90% feldspars and quartz with low density elements and more open structures, will have a low density. The gabbro, S5, contains almost 50% of each type of mineral and would therefore be expected to have a density somewhere between those of S1 and S4.

(d) The feldspar crystals were able to grow unimpeded from the magma, and so they formed good crystal shapes. The quartz was the last mineral to crystallize and had to fill in the spaces left between the other minerals and so is irregular in shape.

(e) All three samples are coarse grained, which implies that they all cooled and crystallized slowly. This would happen at considerable depth below the Earth's surface.

SAQ 2 (a) The overall colour of the rock is very dark, which suggests it might have a high proportion of ferromagnesian minerals.

(b) As ferromagnesian minerals all form at relatively high temperatures, it is likely that this rock formed from a magma at a high temperature.

(c) As the rock is fine grained, it must have cooled and crystallized rapidly, with many crystals growing at once in the magma as it chilled below their crystallization temperatures.

(d) The density of S3 should be similar to that of S5, the gabbro.

It should be clear from this exercise that gabbro (S5) and basalt (S3) have quite a lot in common: their dark colours suggest that both contain a substantial proportion of ferromagnesian minerals. Thus, although we cannot see the individual minerals in the basalt, we can deduce that it has a broadly similar mineralogy to the gabbro. Since the gabbro is coarse grained, and the basalt fine grained, they could have been derived from the same type of magma in which the gabbro cooled and crystallized slowly, and the basalt cooled and crystallized rapidly.

SAQ 3

A Incorrect. Although it is correct that the crust contains a relatively high proportion of radioactive elements, these cannot account for high temperatures in the mantle and core.

B Incorrect. Clearly, the amounts of radioactive elements present today are only a fraction of what they were initially.

C Probably correct. The heat produced inside the Earth today from radioactive sources is marginally inadequate to account for observed surface heat flow. Some primordial heat may still contribute.

SAQ 4 The components removed are clearly SiO_2 , Al_2O_3 and CaO . But, since the andesitic magma has very nearly the same SiO_2 content, there will be only a trivial depletion of SiO_2 in the residual liquid. Al_2O_3 , by contrast, will be much more heavily depleted, and CaO rather less so. If the plagioclase contained *less* CaO than the 8% present in andesitic magma then, of course, the residual liquid would be relatively *enriched* in CaO .

SAQ 5 Only G and I apply only to volcanic rocks, all the others could apply to plutonic rocks. Refer to the sets of criteria listed in Sections 3.2 and 3.4.

SAQ 6 Each kilometre of destructive plate margin erupts $2.5 \times 10^{-5} \text{ km}^3$ of magma each year. The total world production is therefore 50 000 times greater than this, i.e. 1.25 km^3 . But only 80% of this is *new*, derived from the mantle. Therefore, the volume of new continental crust generated each year is $(1.25 \times 80)/100 \text{ km}^3$, i.e. $1 \text{ km}^3 \text{ yr}^{-1}$.

SAQ 7 It must originate from partial melting of pre-existing continental rocks overlying the subduction zone. These may be andesites or granites, or sediments that have been formed at an older plate margin.

SAQ 8 Steady removal of plagioclase and pyroxene crystals from a magma of initially basaltic or andesitic composition will yield a residual liquid of increasingly granitic composition.

SAQ 9 Granites may be formed in small volumes at *constructive* plate margins by extreme *fractional* crystallization of *basaltic* magmas, but they are formed in much larger volumes at *destructive* plate margins, where progressive episodes of magmatism lead to the evolution of progressively more *siliceous* liquids. Heat carried up into the crust may also cause partial *melting* in the upper crust, thus contributing further to the volume of granitic magma.

SAQ 10 Quartz pebbles are unlikely to have come from the quartz veins of Glaslyn lake because of the way in which lakes act as sediment traps. No fewer than four lakes intervene in the river course between the observed quartz veins and locality 6. These quartz pebbles are much more likely to have come from similar veins in rocks that are much closer to locality 6.

SAQ 11 Gabbro. The granite is composed mainly of white potassium-feldspar, quartz and biotite mica, which are low-temperature, stable silicates and, therefore, are fairly resistant to chemical weathering. The gabbro is composed mainly of calcium-rich plagioclase feldspar and pyroxene, which are high-temperature, relatively unstable silicates and, therefore, are not very resistant. You would expect from this that the gabbro would decompose more rapidly than the granite.

SAQ 12 (a) Basalts contain 30% or more ferromagnesian minerals and, when these weather, iron oxides are left as residual products. These form the brown surface colouration of weathered basalt outcrops.

(b) The granite consists of feldspars, micas and quartz. The feldspar will weather to leave residual *clay minerals*; the micas will also weather to leave *clay minerals* if chemical weathering is prolonged; quartz is not attacked and remains as a residual mineral. So the residual products are most likely to be clay minerals and quartz, and possibly some mica that has not decomposed.

(c) The gabbro consists of plagioclase feldspar and pyroxene with, perhaps, a little olivine. The feldspars will weather to clay minerals and the ferromagnesian minerals will completely decompose, leaving only an iron oxide residue. So the residual products are most likely to be clay minerals and iron oxide.

SAQ 13 (a) Using Figure 13 you would expect to find particles with maximum grain diameters within the range 10 mm (moveable at a current speed of 0.3 m s^{-1}) to about 0.15 mm (moveable at a current speed of 0.02 m s^{-1}).

(b) As the drop in current speed is rapid, we should expect all material with grain diameters within the size range 0.15 mm to 10 mm to be deposited. As the range of grain sizes is quite large, we should expect the sediment to be fairly poorly sorted.

SAQ 14 (a) You should look for the largest grain size present because this must represent *at least* the largest grain size the transporting medium could carry. In some specimens there may be small pebbles of between 0.5 and 1 cm in diameter indicating a moderately high-energy environment.

(b) Except for larger fragments, such as small pebbles, which become rounded quite rapidly, most of the quartz grains still show angular and irregular surfaces—they are far from well rounded. This suggests that they cannot have been transported for long periods. As chemical weathering must continue during transportation, this also suggests that the transportation time could not have been extensive, otherwise the milky-white feldspar grains would have decomposed.

SAQ 15 Ca^{2+} and HCO_3^- are removed from seawater by calcareous marine organisms to form calcareous shells. There are also siliceous micro-organisms, which remove $(\text{SiO}_4)^{4-}$ from seawater. This means that the *relative* proportion of Na^+ in seawater must be much greater than that in river water and, consequently, the concentration of Na^+ in seawater as a percentage of the total dissolved ionic species must also increase.

SAQ 16 (a) Incorrect. Rocks are more easily deformed at high temperatures, so they are *weaker*.

(b) Correct. Although the fracture patterns are different in each case, rocks do break under either compressive or tensional stresses, especially at low temperatures.

(c) Correct. If this were not the case, structures built of stone, such as bridges and cathedrals would not be

capable of supporting themselves for centuries, and would soon creep to destruction!

(d) Incorrect. An increase in temperature causes a marked increase in the rate of creep.

(e) Incorrect. Rocks are much stronger in compression. Again, if this were not the case it would be difficult to make buildings stay up, as almost all the stresses in a building are compressive ones due to the weight of the structure.

(f) Incorrect. Joints are almost always the result of tensional stresses.

SAQ 17 At low temperatures, rocks are likely to be brittle and to fracture, so *thrusts* would be produced, the compressive force thrusting one block over another. At higher temperatures, rocks become ductile, so more plastic behaviour is likely, so *folds* would be produced.

SAQ 18 55 km.

Since the surface temperature is 20°C , the temperature must increase by 830°C to melt the rock. Since the thermal gradient is 15°C km^{-1} , this requires a depth of $830^\circ\text{C}/15^\circ\text{C km}^{-1} \approx 55 \text{ km}$.

SAQ 19 1B or C; 2D; 3A; 4B.

1 In order to partially melt sediment, high temperatures of about 700°C are required (Figure 4). Contact metamorphism (A) and low-grade regional metamorphism (D) could not possibly fulfil these conditions. A thermal gradient of 28°C km^{-1} at 30 km gives a temperature of around 800°C (Figure 25), which is above the melting/crystallization temperature of granitic composition rocks. The conditions of very high-grade metamorphism are also suitable for melting sediments.

2 Phyllite is a result of low-grade regional metamorphism. The metamorphism is of a slightly higher grade than that which produces slate, but it is lower than that which produces schist or gneiss. Play the AV tape again if you have difficulty with this.

3 Hornfels results from the 'baking' of rocks in contact with hot magma, often plutons. They are therefore characteristic of contact metamorphic zones.

4 Gneiss is one of the highest-grade metamorphic rocks, the last step in the rock cycle before partial melting leads to igneous rocks being formed. Gneisses are formed deep under the mountain ranges of continents, in regions of high-grade regional metamorphism.

SAQ 20 Slate, phyllite, schist, gneiss (see Table 14 and Section 5.4).

Slate is formed at low temperatures and pressures and phyllite at only slightly higher pressure. Schists show extensive mineral growths of biotite aligned at right angles to the direction of pressure. Higher temperatures and pressures are involved than for slate or phyllite. Gneiss represents the last step before re-melting to a granite, and so is the highest grade of these rocks.

SAQ 21 The main criterion is the *texture* of a rock: the texture of the metamorphic rock is crystalline (interlocking grains), whereas the texture of the sedimentary rock is fragmental.

Layering in metamorphic rocks is due to either

- (i) mineral growth at right angles to the direction of pressure (alignment of minerals), or
- (ii) segregation of mineral growth into layers of different composition—ferromagnesian minerals in dark bands, quartz and feldspar in light bands.

Sedimentary layering is a time-dependent feature. The layers are formed at different times, those higher in a sequence being younger than those found lower down. Different layers may be a function of grain size (i.e. of fluctuations in the energy of the transport medium) or of mineralogy (minerals may accumulate in layers according to their density).

SAQ 22 (a) Hawaiian volcanoes are less dangerous than Andean ones mainly because of their composition (Sections 3.2 and 3.5). Hawaiian volcanoes erupt low-viscosity, 'runny' basalts, which rarely solidify within the vents. Andean material is more siliceous and viscous; andesites are formed by the partial melting at the junction of the base of the continent and the down-moving oceanic lithosphere. They have a lower melting (or solidifying) temperature, so vents become blocked with solid material and eventually explosive volcanism takes place.

(b) Sediments are formed by the weathering and erosion of existing rocks (Section 4). The material is moved from the source region by water, wind or ice and deposited as sediment in, for example, oceans, river basins and deltas.

(c) Mud is changed to slate by cementation, lithification and then mild metamorphism (Section 5.4).

(d) The processes in (a)–(c) are linked through the rock cycle, in which rocks are created, weathered and eroded, re-deposited and sometimes re-melted to form new igneous rocks (Figures 2 and 26).

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PLATE 1 The sea cliffs at St Ann's Head, near Dale, on the Pembrokeshire (Dyfed) coast, south-west Wales. They are formed of sedimentary rocks and the layers are clearly visible.



PLATE 2 Arthur's Seat, Edinburgh, Scotland. The prominent saddle-shaped feature in the centre of the picture is the solidified vent of an ancient volcano and is formed of volcanic rocks, which are more resistant than the surrounding material to erosion.

PLATE 3 Giant's Causeway, County Antrim, Northern Ireland. Here extrusive basalt, similar to sample S3 in the Kit, shows characteristic polygonal cracks formed during the cooling of a lava flow (Section 5.2). The cooling cracks can be seen in plan view in the foreground and in section in the middle ground, where it is clear that they extend down into the lava flow.



PLATE 4 The granite cliffs of Land's End, Cornwall. These cliffs are around 70 metres high and formed of granite similar to sample S1 in the Kit (though here they are weathered, and covered by moss and lichens). The roughly rectangular pattern of cracks visible is the *joint pattern* of the rock (Section 5.2).



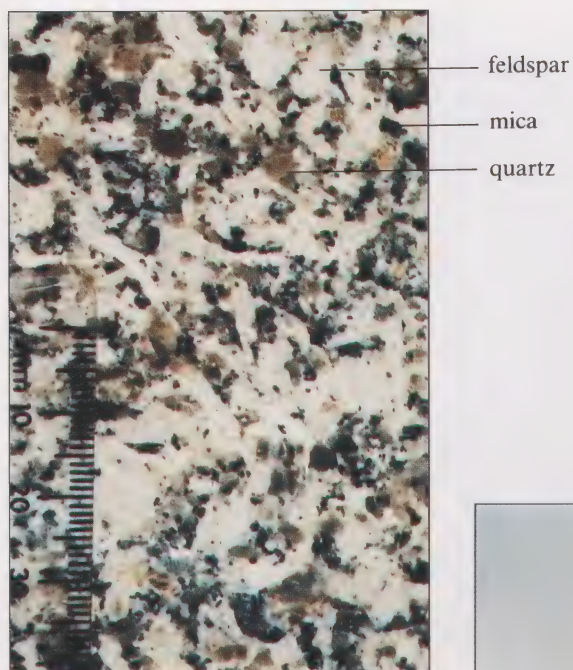


PLATE 6 Polished granite, with quartz, feldspar and mica labelled. Feldspar can be white, pink or shades of grey. Quartz in rocks is usually clear pale grey.



PLATE 5 (a) Specimen of coloured quartz (amethyst) from Brazil. (b) A specimen of quartz in which the crystals have grown in a void, allowing their shapes to develop unhindered.



PLATE 7 A large mica crystal. The flat shiny surface is a cleavage plane and there are many similar parallel planes within the crystal. Crystals of mica like this one are sometimes called 'books', because the edges resemble the closed pages of a book.



PLATE 8 A calcite (CaCO_3) crystal, with three cleavage directions parallel to the crystal faces.

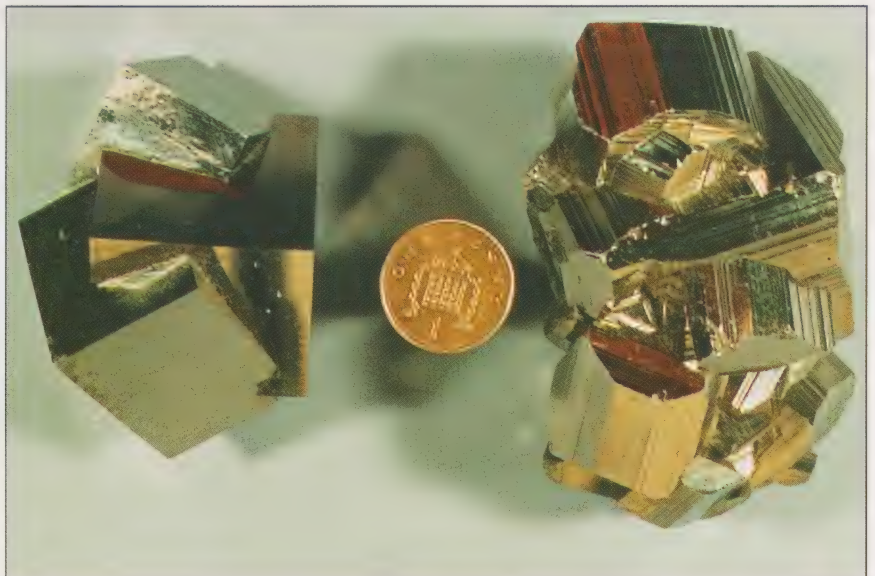


PLATE 9 Crystals of pyrite, which has metallic lustre. The combination of colour and lustre led to this mineral's common name, 'fool's gold'. The mineral forms almost perfect cubes, often intergrown, the faces of which often show parallel lines or *striations*. These are visible on the faces of the larger sample in the photograph.



PLATE 10 (a) A fissure eruption in Iceland in October 1980. The fissure from which the lava issues is several kilometres long and there is 'fire fountaining' along most of its length. The lava then flows away from the fissure, downhill in 'rivers of fire'. (b) A more explosive type of eruption in Hawaii in 1960. Rapid release of gas from hot fluid magma causes fire fountaining above the crater of volcano Kilauea.

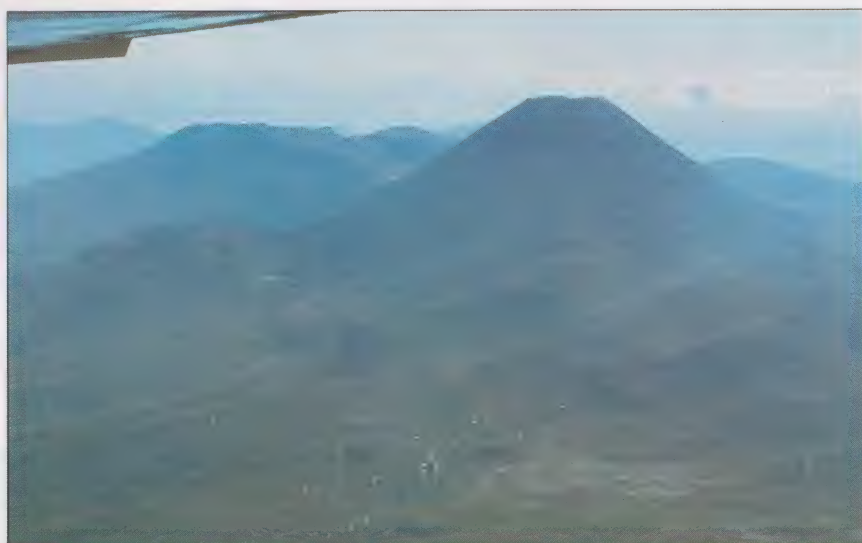


PLATE 11 Two andesite volcanoes in the Tongariro National Park area of North Island, New Zealand. Ngauruhoe, the conical structure in the foreground, has a single crater and is one of the youngest (2 500 years old) and most active volcanoes in New Zealand. The Ruapehu massif, behind, is a complex volcano. It consists of several formerly active craters beneath a cone structure (like Ngauruhoe), the top of which was blown off about a quarter of a million years ago.

PLATE 12 The small-scale cross-bedding characteristic of water-borne deposits. Compare the scale of the cross-bedding here with that shown in Plates 13a and b.

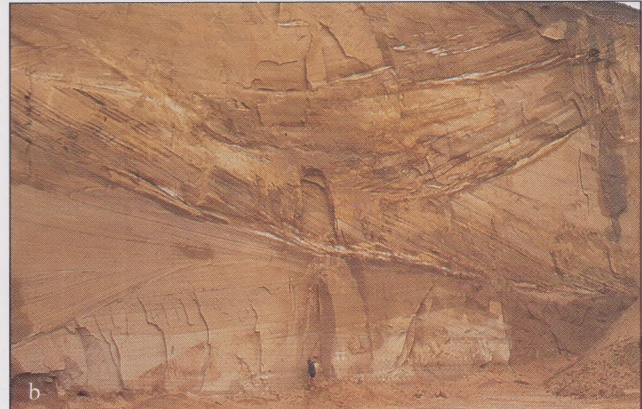
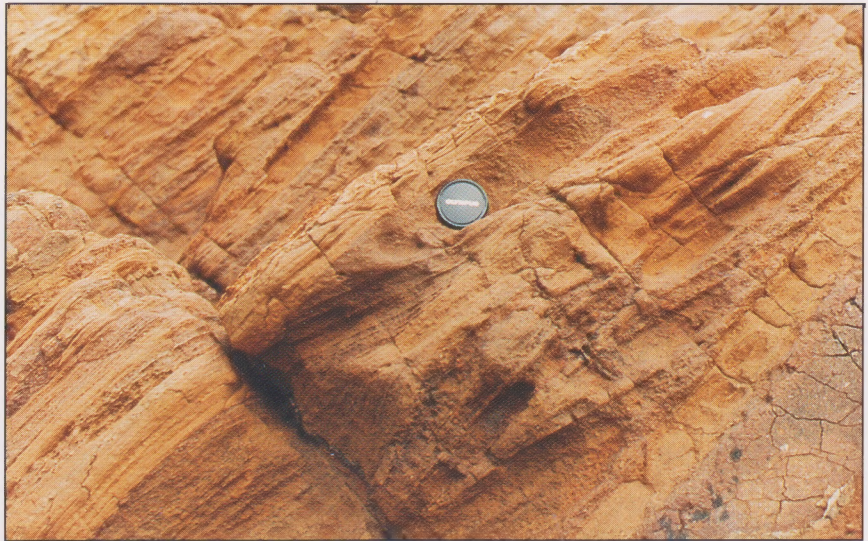


PLATE 13 (a) The sand dunes of Death Valley, California. The surface of the dunes is covered with small ripples caused by the wind. (b) A cross-section of an ancient sand dune, exposed in a quarry in the North of England. Dunes like this in the geological record provide evidence that Britain experienced desert conditions in the past, when it was much closer to the equator. You will see that the scale of the cross-bedding shown here is far larger than that in Plate 12 — it is in the order of metres, rather than centimetres (note the relative size of the person).

PLATE 14 Material deposited when the ice of a glacier melts is called *moraine*. This photograph shows the moraine at the end of the Aletsch Glacier in Switzerland. The characteristic of ice-borne deposits is that they are very poorly sorted, a jumble of different sized particles from very large boulders to extremely fine clay. Additionally, the particles are very angular, as they have not been rounded by multiple collisions with other particles during transport.





PLATE 15 Map of part of North Wales, near Snowdon (for use with TV 27 'From Snowdon to the sea'). The localities visited during the programme are numbered 1–8.

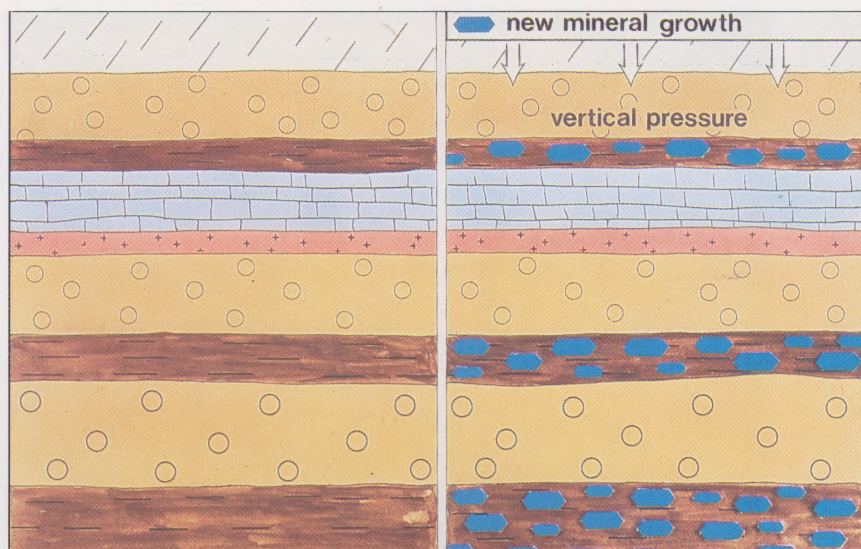


PLATE 16 Crystal growth along bedding planes, where pressure is due to the weight of overlying rocks.



PLATE 17 An outcrop in Wales, showing cleavage planes cutting across the original bedding.

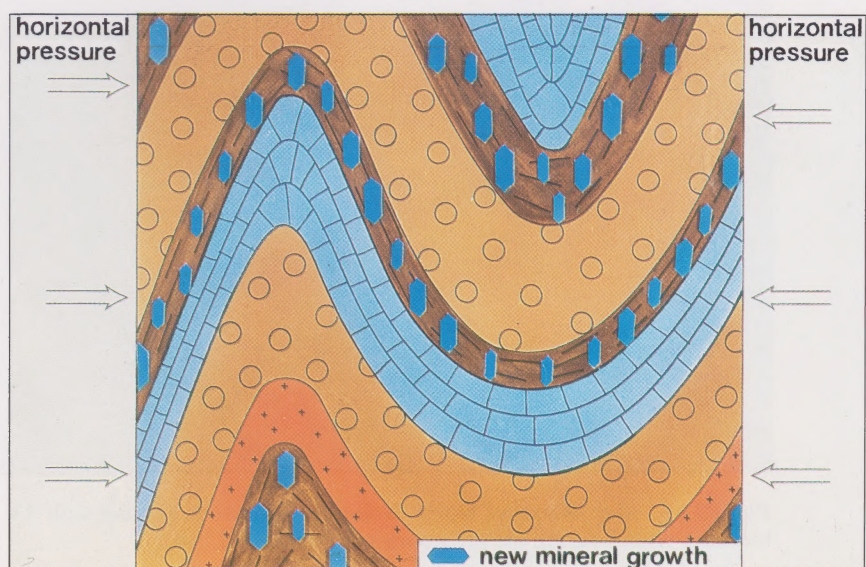


PLATE 18 Horizontal, or sideways, pressure showing how cleavage develops.

PLATE 19 Outcrop of schist, folded into complex shapes, in Norway.



PLATE 20 (a) Glaciated slab of gneiss. (b) Sample of gneiss in the Kit, showing light minerals (largely feldspar and some quartz) and dark minerals (iron-rich, biotite mica) in bands.

